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Journal of Sedimentary Petrology

A Publication of the Society of Economic Paleontologists and Mineralogists
a Division of
The American Association of Petroleum Geologists

« CONTENTS »

Pennsylvanian sedimentation in Northeastern Socorro County, New Mexico ..... *Raymond Sidwell and G. Frederick Warn* 3
Gamma ferric oxide in sediments ..... *H. W. van der Marel* 12
Sea lions as geological agents ..... *C. A. Fleming* 22
An unusual stalactite from Saltville, Virginia ..... *F. G. Snyder* 26
The problem of dolomite formation considered in the light of research on dolomites in the Sonnwendmountains (Tirol) ..... *Robert Weynschenk* 28
A comparison chart for visual percentage estimation.. *Robert L. Folk* 32
Some occurrences of supergene iron sulphides in relation to their environments of deposition .... *A. B. Edwards and G. Baker* 34
Longitudinal ripple marks in mud and sand.. *L. M. J. U. van Straaten* 47

REVIEWS

Tj. H. van Andel. Provenance, transport and deposition of Rhine sediments ..... *Lincoln Dryden* 55
F. H. Hatch, A. K. Wells and M. K. Wells. Textbook of Petrology, vol. one, The Petrology of the Igneous Rocks ..... *Carleton A. Chapman* 58
F. H. Hatch and R. H. Rastall. Textbook of Petrology, vol. two, The Petrology of the Sedimentary Rocks. *J. L. Hough* 59

ANNOUNCEMENT

Creation of an Association of French sedimentologists 60

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Articles submitted for publication to the *Journal of Sedimentary Petrology* should have a prefatory abstract and sufficient postage should be sent by the author to pay the return of the manuscript as it has been found that many articles require more or less revision before publication.

Persons who are not members of the Society of Economic Paleontologists and Mineralogists or are not subscribers to the *Journal of Sedimentary Petrology* but who desire to publish an article in the *Journal*, should have some member of the Society of Economic Paleontologists and Mineralogists or some member of the American Association of Petroleum Geologists sponsor the article. Members of these Societies are given preference in publication.

Among papers to appear in forthcoming issues of the *Journal* are the following:

Occurrence and lithologic associations of evaporites in the United States. By W. C. Krumbein.

Pisoliths, ooliths, and calcareous growths in limestone caves at Port Campbell, Victoria, Australia. By George Baker and A. C. Frostick.

Beach sands of the Mediterranean coast of Egypt. By Mohamed E. Hilmy.

A new sedimentary petrological interpretation of the results reported by the Swedish "Albatross" deep-sea expedition—1947-1948. By Robert Weynschenk.

Trends in literature on sedimentology. By K. O. Emery.

Further discussion of "What is sedimentology?" By E. Alan Lohse.

JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 21, NO. 1, PP. 3–11
FIG. 1, PLS. I–II, MARCH, 1951

PENNSYLVANIAN SEDIMENTATION IN NORTHEASTERN SOCORRO COUNTY, NEW MEXICO

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Texas Technological College, Lubbock, Texas

ABSTRACT

Pennsylvanian sediments in northeastern Socorro County, New Mexico, in order of abundance, consist of shales, limestones, and sandstones and attain a thickness of about 740 feet. Insoluble residues, organic remains, and thin sections were utilized to determine characteristics of sediments, metasomatic changes, and environment of deposition. Metasomatic changes include: calcification, aragonite (?) to calcite; dolomitization, calcite to dolomite; silicification, calcite and dolomite to chalcedony; and alteration of undetermined minerals to illite. Sedimentation, in a relatively unstable, shallow geosyncline, was influenced by four positive areas of low to medium relief.

INTRODUCTION

In northeastern Socorro County, New Mexico, almost flat-lying Pennsylvanian rocks are contiguous with older highly tilted rocks of the Manzano and Los Pinos mountains. The exposed sediments attain a thickness of about 740 feet and consist chiefly of limestones alternating with shales and arenaceous materials. This normal sequence of sedimentation was interrupted at two widely separated intervals by the deposition of arkosic materials.

The areal geology of Socorro and adjoining counties was mapped by Wilpolt and MacAlpin (1946) who subdivided Pennsylvanian rocks into two formations —Sandia, basal, and Madera, upper. The measured section herein described is well exposed in Abo Canyon. However, the Sandia which normally overlies the pre-Cambrian, was downfaulted and concealed during Tertiary orogenic movements.

Materials collected from each lithologic unit were processed for insoluble residues. Heavy minerals were removed from arenaceous sediments by the bromoform method. For microscopic study thin sections were prepared from specimens of limestones and sandstones. Microphotographs were used to supplement the direct observation of the thin sections. The study is primarily one of sedimentary conditions during Madera time. Consideration is given to sedimentary sources, transportation, depositional conditions, tectonic movements, and post-depositional alteration of sediments.

STRATIGRAPHY

The exposed Madera sediments in respect to abundance are: 49 per cent

FIG. 1.—Map of New Mexico showing general location of area.

shales; 38 per cent limestones; 9 per cent sandstones; and about 4 per cent arkosic materials and cherts. The shales, interbedded with thin sandy layers, range in color from dark gray to greenish-gray and attain a maximum unit thickness of about 50 feet. The calcium carbonate content, as much as 30 per cent, is high in the dark gray shales, especially those containing calcareous marine organisms, and low in the greenish-gray shales where organisms are rare. The laminations, ranging from paper thickness to more than a foot, are marked by mica flakes arranged with the long axes parallel to planes of separation. Organic remains, especially sponge spicules, coral, bryozoa, shells of brachiopods and tests of foraminifera, are numerous in the lower part of the section, whereas plant fragments are about the only fossils represented in the upper shales. The shales in the upper part of the section are highly indurated and contain an abundance of seritic clays. In some units gypsum is the chief binding material. Some of the thick shale layers, especially those with an abundance of silty material, show minor flexures, probably resulting from compaction. The contact between the shales and coquinoid limestones is marked by a rather abrupt change from limestones to silty clays, and by the decrease in abundance of organic shells. However, the upper contact between the shales and overlying limestones is gradational, and is characterized by calcareous nodules and lenses.

The normal sedimentation sequence in the extensive units of calcareous materials includes: a basal nodular layer; a medial, black to dark gray limestone; and upper biostromal beds. The basal limestones, a gradational zone in the shale-limestone contact, occur as nodular and thin bedded calcareous layers attaining a total thickness of 2 to 6 feet. The nodules are circular to elongated, marked by an irregular surface and occasionally by an interior cavity lined with well developed quartz crystals. Thin bedded, dark gray, moderately fossiliferous limestone layers are separated by thin shales. The shale partings are 1 to 3 inches thick.

The medial, black limestones constitute about 75 per cent of the calcareous unit. They are dense, break with conchoidal fracture, contain secondary chert nodules, and commonly grade upward into lighter colored beds which are slightly oolitic. The texture of the dark limestones as determined from thin sections is porphyroblastic with porphyroblasts of dolomite and a groundmass of small calcite particles. The insoluble residues, from 3 to 7 per cent of the rock, consist chiefly of black clay, a few mica flakes, and chert.

The coquinoid, flaggy limestones, from 3 to 10 feet thick, contain an abundance of organic remains consisting chiefly of fragments of echinoderms, bryozoa, coral, shells of brachiopods, and tests of fusulinids. The fossils are poorly assorted, do not show evidence of rolling and are embedded in a fine grained matrix, chiefly clay and silt. The flaggy limestones grade upward into gray micaceous shales associated with calcium sulfate. The depositional environment was apparently unfavorable to organic activity while these gray shales and gypsiferous materials were accumulating.

Rocks composed of sand size particles constitute less than 9 per cent of the Pennsylvanian sediments and occur as thin layers interbedded with shales. The sands are gray to greenish-gray, moderately well sorted, and contain fragments of plants. The carbonate content, from 2 to 30 per cent, consists of clastic limestone particles and calcite cement. Mineral constituents of the sands in order of abundance are: quartz and chert; limestone fragments; mica; tourmaline; iron oxides; garnet; epidote; anhydrite; and zircon. The sand particles, chiefly sub-angular, are of medium texture and possess characteristics suggestive of two source materials, metamorphic and igneous. Limestone fragments were derived from underlying beds. Chert and secondary silica occur in irregular detrital masses, as secondary growth replacing

calcareous material, as overgrowths on quartz grains, and as banded fracture fills. Many of the quartz grains are pitted and fractured. Thin section studies show finely granular calcareous material occupying these openings. Gray to black silt and clay particles fill much of the pore space between the grains. Some of the sands in upper layers contain brown stains which in thin section appear as marine plant fragments preserved in hydrous iron oxide. Mica, chiefly muscovite, is the chief constituent of some sands.

Two zones of arkosic materials disrupt the limestone-shale sequence. One occurs about midway in the section, and a second in the upper part marking the close of marine sedimentation. Materials in the lower arkosic zone, chiefly angular particles of quartz and feldspar and hexagonal mica flakes, attain a thickness of about 16 feet. The deposit is characterized by poor sorting, grayish color, clay and silt matrix and predominance of coarse sand sizes. The feldspars, chiefly orthoclase, constitute about 10 per cent of the rock and some of the grains are altered to kaolinite, Carbonate content varies from 16 to 28 per cent and occurs chiefly as a cementing material.

The arkoses in the upper part of the section were associated with crustal movement of sufficient intensity to close the geosyncline. Arkosic materials of marine origin first appear in the upper limestone unit and grade upward into normal non-marine arkoses. The arkosic materials, primarily of pebble sizes, are dispersed through about 10 feet of the limestone and consist chiefly of quartz, feldspar and mica flakes. Numerous limestone fragments occur in the upper layers, making an arkosic limestone with about 75 per cent carbonate content. The feldspars embedded in the limestone are practically unaltered. The overlying normal arkoses constitute the transition beds in the Pennsylvanian-Permian contact. They are characterized by reddish color, poor sorting, high clay content, and angular quartz grains, some of the latter attaining pebble sizes.

Thin sections of arkosic materials show a minor amount of secondary quartz and of chert with clay inclusions in addition to granular calcium carbonate and calcite. The limestones associated with arkoses contain fossil fragments and spherulites.

DIAGENESIS

The term diagenesis is used to include processes such as metasomatism, authigenesis and cementation. The metasomatic process involves replacement of minerals and appears to follow these sequences: aragonite (?) to calcite; calcite to dolomite; and calcite or dolomite to chalcedony. Other metasomatic changes have altered clay minerals to sericite. The authigenetic process includes the enlargement of detrital grains by overgrowths and the development of secondary minerals. Cementation refers primarily to chemically introduced binding materials such as calcium carbonate, calcium sulfate, and quartz.

The limestones at the time of deposition were an aggregation of small calcite particles, calcareous remains of marine forms, organic muds, and variable amounts of calcareous muds, mica flakes and subrounded quartz grains. Some of the effects of replacement in the limestones include: an increase in complexity of mineral composition, a change from mosaic to porphyroblastic texture, a variable age diversity of the minerals, and the removal of original textures and structures.

Aragonite, one of the least stable of the carbonates, was no doubt a constituent of oolites and some of the organic remains, especially algal growths, tests of foraminifera, coral, bryozoa, and shells of molluscs. Most of the aragonite (?) was changed to calcite and during the diagenetic process the host mineral, except for typical concentric, laminar and radial structures, was destroyed. (See fig. 2, pl. I.)

Calcite occurs as an original mosaic of small granular particles, as a constitu-

ent of some marine shells, and as a replacement mineral after aragonite (?). The metacrysts of calcite are distinguished from the original calcite by the large, clear, irregular, well outlined crystals, by the well developed cleavage, and by the inclusions of small clastic particles.

The change from calcite to dolomite is concentrated in the light gray, thin to medium bedded, fossiliferous limestones. The dolomite, chiefly as large irregular crystals, occasionally as rhombs, supersedes most of the metasomatic calcite, especially in the interior of shells and oolites, and some of the original calcite. The sequence from the interiors of shells and oolites outward includes: a large nuclear crystal surrounded by two or more rows of smaller particles; a crystal-

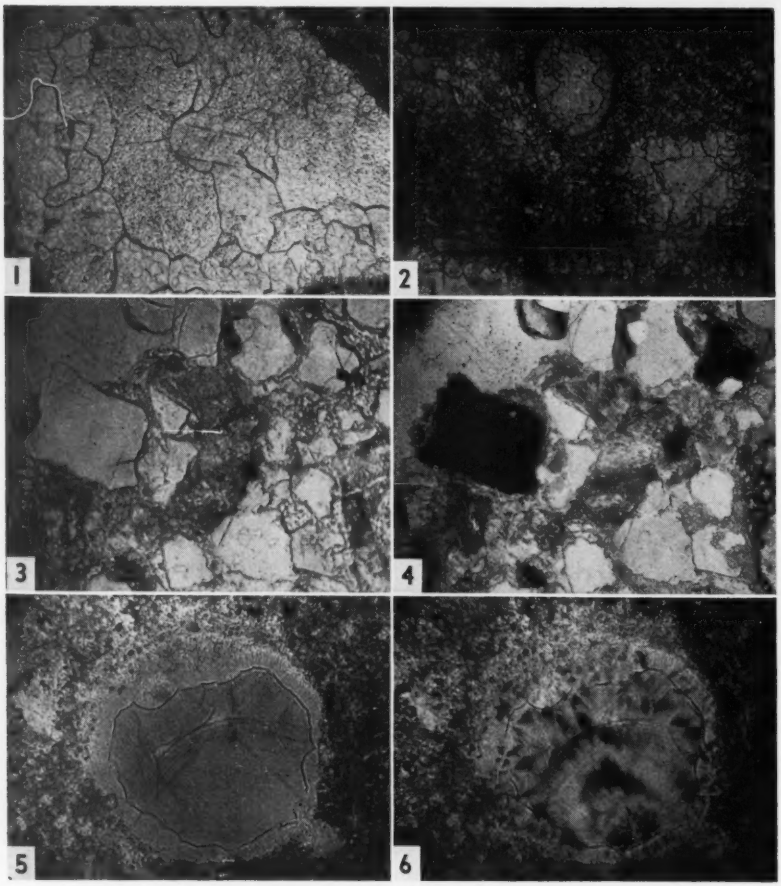

PLATE I

line calcite zone; and a concentric outer layer of calcareous material. (See fig. 6, pl. II.) Dolomite in the original calcite mosaic occurs as large, irregular disseminated metacrysts and as crystal aggregates. The sequence of materials in aggregates follows the pattern outlined for the interior of shells.

The dolomite metacrysts in the aggregates present an age diversity in respect to time of growth. The sequence of growth appears to be in the following order: first a nuclear crystal succeeded by rows, two or more, enclosing the nucleus, continued growth outward replacing calcite and/or calcareous muds. (See fig. 2, pl. I.) The pattern is marked by a decrease in particle size from center outward. Each successive metacryst occupies re-entrant cavities some of which are sharply concave or shell-like, and others are smoothly concave. According to Lindgren (1928) the concave boundary suggests replacement. The absence of voids, collapse cavities, and crystal crowding negate the possibility of post-lithification leaching and replacement.

Silicification, the change of calcite or dolomite to chert and chalcedony, is concentrated in the dark gray to black, medium to thick bedded limestones. The siliceous material is present as nodules, as disseminated grains and as spherical masses and stringers of nearly pure chalcedony. The nodules are dark gray to black, of varied shapes and sizes, and are structureless. Their occurrence in limestone is devoid of pattern. The contact between the nodules and host rock is irregular and contains a thin layer of nearly black clay. The disseminated chert particles are irregular in outline and consist primarily of minute radiating fibers.

EXPLANATION OF PLATE I

FIG. 1.—Crystalline veinlet—Replacement of fine calcareous muds in a dark, dense limestone by large crystals of dolomite and probably some calcite. Small dark areas within the crystals are inclusions of calcareous muds, typical of dolomite replacement. Dolomite rhombs are suggested in the center. The dark crystal boundaries are made of very fine granular mud or organic matter. The larger crystals in turn contain smaller masses which may represent dolomite replacement of calcite. Magnified about 190 diameters.

FIG. 2.—Dolomite replacement of a finely crystalline calcite mosaic in a dark gray dense limestone. The same dark crystal margins obtain. The rounded object is an oolite with an outer dark radial layer suggestive of aragonite structure. The suture effect is a dolomite boundary while the central portion represents calcite (?) replacement. Dolomitization was in progress from the outer part of the oolite toward the center. The nucleus has been entirely replaced, and concentric layers typical of the original oolitic structure are lacking. The same order of replacement is indicated in the crystal mass to the SE of the oolite where a nuclear crystal appears in the center. The dark wire-like object NW of the oolite is a section through an arenaceous foraminifera. Magnified about 125 diameters.

FIG. 3.—Gray sandy limestone showing fractured clastic calcite and quartz embedded in finely divided calcite and calcareous mud. Calcification and silicification are indicated by overgrowths on many of the angular fragments. Inclusions and dust rings are visible in most of these fragments. Silica forms the cementing material in the NW corner. Magnified about 125 diameters.

FIG. 4.—The same slide area under crossed nicols. In the NE corner the darkened fragment shows a light border representing addition of silica. The same type of growth is suggested along part of the border of the large dark fragment. In the SE corner calcite (?) replacement of calcareous mud is indicated. Magnified about 125 diameters.

FIG. 5.—Replacement of calcareous and organic muds by a chert nodule in a dark gray, dense, cherty limestone. Concentric bands and radial lines within the dark boundary show the growth patterns of chalcedonic quartz. The outer lighter border is granular silica which was actively growing outward into the darker granular calcareous-organic mud. The light isolated patches surrounding the nodule may represent siliceous impurities in the original sediments. Several dark inclusions show near the center of the nodule.

FIG. 6.—The same slide area under crossed nicols. Concentric and radial growth lines are distinct. Radial lines are continuous into the granular silica but concentric banding is lacking. The dark coloform band is opaline silica. Magnified about 125 diameters.

The spherical masses of nearly pure chalcedony are composed of microcrystalline materials and are marked by three or more successive zones of growth. The normal sequence of growth consists of an infill of radiating fibrous crystals, a thin dark colored coloform band, and an outer zone with indication of both radial and granular structure. The outer zone extends irregularly into the surrounding calcareous material. (See figs. 6 and 7, pl. I.)

Some of the Pennsylvanian shales contain a mineral similar to illite which was named and described by Grim, Bray and Bradley (1937). The name was suggested "as a general term for the clay mineral constituent of argillaceous sedi-

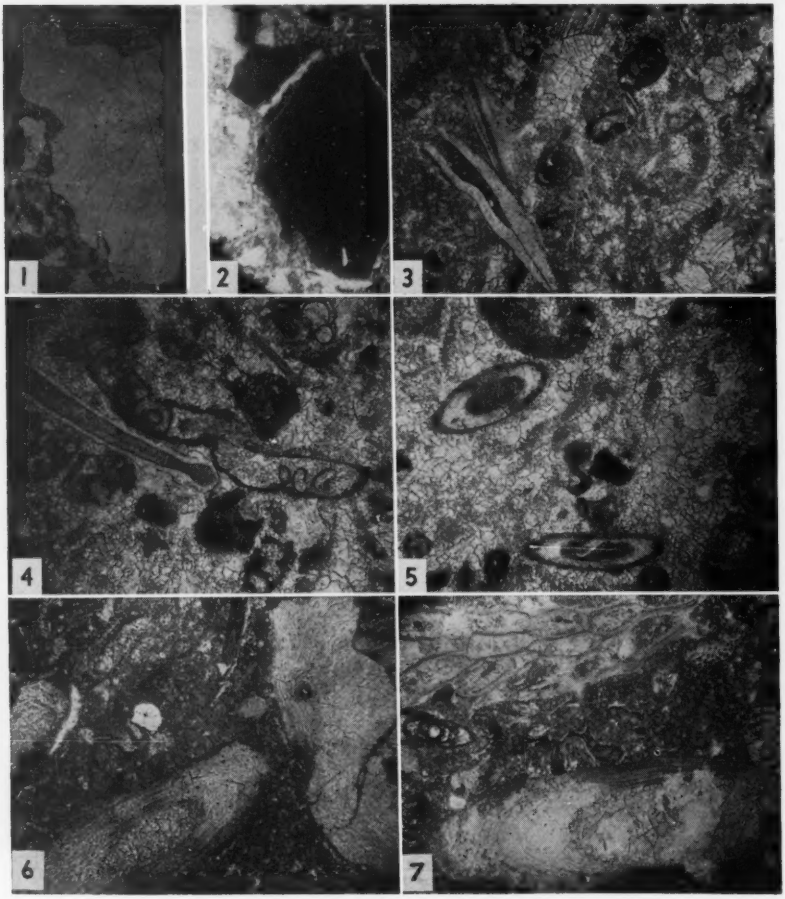

PLATE II

ments belonging to the mica group." Illite in the shales is an impure secondary mica with a low potassium content. It occurs as shred-like particles and as minute flakes resembling sericite. These sericitic shales are indurated and give poor separation in the residue. They are almost devoid of calcareous organic remains.

Shales containing illite occur in units 30 feet or more in thickness. Spectograph records of the illite show a higher potassium content than is found in the average shale. A single shell fragment replaced entirely by illite suggests a metasomatic origin for this mineral. The sericitic materials are associated with partly carbonized plant remains, with little altered biotite flakes and with a high gypsum content. Most of the gypsum is present as a cementing material.

The authigenic process involves the development of overgrowths on quartz and calcite and the formation of new

EXPLANATION OF PLATE II

FIG. 1.—Fractured chert fragment showing silicification. Silica replacement of calcareous mud has recently occurred along the edges of the chert fragments. Dark inclusions are scattered through the chert. The dark lines represent fractures outlined by inclusions. See below. Magnified about 300 diameters.

FIG. 2.—The same fragments under crossed nicols. The light areas represent secondary silica. Notice the veinlet effects of silica additions across the large fragment and the light silica border around the small fragment immediately above. Magnified about 300 diameters.

FIG. 3.—Microcoquina in a dark, dense limestone. The groundmass is finely granular $CaCO_3$ with calcareous and organic muds. In the upper center, cleavage planes of the carbonates are distinct. To the right is a *Lagena* type foraminifera in inverted position. In the center is a section through a quinqueloculine foraminifera. Fragments of smaller agglutinated types appear in the upper left. The larger organic fragments are parts of shells. Magnified about 300 diameters.

FIG. 4.—Matrix of coarser carbonate crystals enclosing fine calcite particles and organic remains. The elongate form extending from the upper left is *Earlandia*, an agglutinated foraminifera. The dark bulbular end is the subglobular proloculum, extending into a long tubular second chamber. The chamber walls are light with a heavy, dark inner border and a very thin, irregular outer border. Another agglutinated form with spherical chambers, dark inner wall border, and thin irregular outer border appears near the top on the right. Near the bottom on the right is a form suggesting *Endothyranella*. The trail of organic debris from the upper left across the center suggests fecal pellets. Replacement has had little effect on organic structures and the effects of compaction have been slight. Magnified about 300 diameters.

FIG. 5.—Crystalline limestone with a mosaic of calcite partly replaced by dolomite. The oval and elongate oval with dark centers and dark outer borders are fusulinid foraminifera. The interiors are largely filled with mud. The thin concentric lines represent septa which are being replaced by calcite. To the left of the lower fusulinid is a saggittal section, probably through a megalospheric form, with the central proloculum bordered by volution chambers. The large round dark area in the upper center suggests a pisolite with nuclear replacement by calcite. Relatively large dolomite crystals appear in the matrix but have invaded none of the organic structures. Calcite replacement in the forams has started in an intermediate zone and appears to be working into both central and outer zones. On the extreme right is a circular growth of calcareous algae. Magnified about 300 diameters.

FIG. 6.—Dark gray fine grained limestone. The dark background is organic and calcareous mud. In the lower left a fusulinid has had dolomitic replacement of the central zone. The concentric septal layers were being replaced by calcite which was extending into the surrounding muds. Dolomite replacement in the center appears to be following the calcite in an outward direction. The large crystalline mass on the right seems to show dolomite replacement of granular calcite. A similar patch of granular calcite on the far left has not been so affected. The white fragment with small circular patterns at left center of the picture is an echinoid plate fragment. Magnified about 300 diameters.

FIG. 7.—Microcoquina in a gray nodular, sandy limestone. In the upper part are bryozoan fragments and on the right side center is a fragment of a crinoid columnal. On the left side center is an axial section of an immature fusulinid. The central constriction is the chomata and the central light sphere the proloculum. Across the lower part of the picture is a mature fusulinid in which dolomite replacement of calcite has started in the interior. The outward growth of calcite is absorbing the calcareous mud and is breaking through the septal folds to the outside of the test Magnified about 300 diameters.

minerals, especially kaolinite and pyrite. Chalcedony and quartz overgrowths on quartz particles occur chiefly as layers on the grain margins and are commonly marked by discontinuity with the original grain. Some of the overgrowths are separated from the original quartz particles by a chain of clay or calcite inclusions. (See figs. 3 and 4, pl. I; figs. 1 and 2, pl. II.) Overgrowths of calcite, less numerous than in quartz, occur as a film of calcium carbonate on the margins of the grains and as fillings in fractures in calcareous particles. (See figs. 3 and 4, pl. I.)

Some of the feldspars in arkosic materials have been partly altered to kaolinite, while in others the alteration is complete. Kaolinite, a gray opaque mineral, is also an important interstitial material in the arkoses. Pyrite occurs as small cubes in the dark gray to black limestones. It is associated with plant remains that are in a low state of oxidation.

ENVIRONMENTS OF DEPOSITION

Reiche (1949) indicated that early Pennsylvanian sediments in northern New Mexico were laid down on a peneplained surface, one exposed to subaerial processes from pre-Cambrian time to Pennsylvanian time. Sedimentation was influenced by a number of positive areas of which the most conspicuous, according to Read and Wood (1947), include: the Uncompahgre Axis to the northwest; the Sierra Grande Arch to the northeast; and the Pedernal Axis and Joyita Axis to the south. The greatest influence on the deposition of Madera sediments in Socorro County was, because of its proximity, made by the Pedernal Axis.

Madera sediments, the result of long continuous deposition, were laid in a sea probably less than 600 feet deep. Shales and sands, excepting the arkoses, reflect borderlands of low to moderate relief. Thick limestones alternating with shales and sands suggest a depositional basin characterized by long intervals of quiescence alternating with relatively short turbulent phases. Apparently, some of the limestones were deposited in wave and current protected areas where accumulation of organic matter and its incomplete bacterial decay were favored. Slow subsidence of the negative areas is implied by the thick accumulation of materials which record an apparent balance between the rate of sediment accumulation and the rate of sinking.

The positive areas, especially the Pedernal pre-Pennsylvanian land mass, were characterized by a metamorphic-granitic terrane of moderate to low relief, and by comparatively slow continuous uplift. The rate of uplift probably was accelerated prior to and during the deposition of arkosic materials. The Pennsylvanian-Permian uplift was slow and intermittent as indicated by alternating marine and nonmarine beds.

Supporting evidence for the above assumptions includes: 1. The normal sequence of deposition, limestones alternating with arenaceous shales, suggests mildly unstable source areas. 2. The dark gray and black cherty limestones are massive and contain detrital materials, chiefly black clay and mica flakes. They represent a depositional environment where free circulation and complete oxidation were impeded. 3. The normal sequence of sedimentation was disrupted by two zones of arkosic material, one about midway in the section, a second at the top of the section. These materials are poorly sorted and the mineral content consists of angular particles of quartz and feldspar together with micas in a fine grained matrix. The physical character of the arkose deposits suggests: 1. A granite provenance of moderate relief. 2. Moderate but accelerated uplift in the positive areas. 3. A predominantly physical destruction of parent rocks followed by short transportation and rapid deposition. The mineral assemblages, inclusions in quartz, and an abundance of fine grained materials (clays and silts) suggest a metamorphic-granitic terrane of medium to low relief. Alternate marine and non-marine arkosic sands and arkosic

limestones occur at the top of the column. They record a closing phase of intermittent uplift accompanied by oscillatory Pennsylvanian seas.

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GAMMA FERRIC OXIDE IN SEDIMENTS

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ABSTRACT

It was found by X-ray analysis that lepidocrocite (γFeOOH) or limonite (αFeOOH) in the presence of organic matter (humus, amylum, citric acid, oxalic acid, etc.), are converted to γFe_2O_3 = maghemite when heated in the air. The γFe_2O_3 (with an unit cell of 8.39 A) obtained from natural epidocrocite or limonite concretions, which commonly contain small amounts of sodium or managanese in their natural state, can withstand heating for 2 hours at about 600° C.

Furthermore, stabilized γFe_2O_3, having a unit cell of 8.48 A, was obtained by heating ferric citrate containing sodium as an impurity.

In nature γFe_2O_3 is formed by slow reduction and afterwards oxidation of lepidocrocite (γFeOOH) or limonite (αFeOOH) particles (concretions) rich in organic matter. Thus γFe_2O_3 may be a common mineral in sediments (soils) rich in iron and organic matter, such as the brown-red colored peaty soils of the provinces of Groningen and Drenthe in the Netherlands. The reduction—oxidation process may be accelerated which happens when these peats are set on fire by accident or with the purpose to obtain better crops.

INTRODUCTION

Baudisch and Welo (1925) obtained γFe_2O_3, a red-brown iron oxide mineral of the spinel type (XY_2O_4), also called maghemite because of its magnetic properties, by careful heating of synthetic magnetite (Fe_3O_4). The synthesis of magnetite can be performed in two ways; i.e., after Lefort (1852) and Kaufmann (1900–1901) from, respectively, a solution of Fe^{2+} and Fe^{3+} by precipitation with alkali or a solution of Fe^{2+} by oxidation in the presense of alkali. It is further stated by Hahn (1923), Baudisch (1932) and Glemser (1938), that γFe_2O_3 can also be obtained from γFeOOH (lepidocrocite, iron rust) by heating at 200°–400° C. in vacuum. Gamma FeOOH occurs in nature or it may be synthesized according to the methods as given by the above authors; respectively from a solution of $Fe^{2+} + NH_4Cl + Na_2S_2O_3 + KJO_3$, from a solution of Fe^{2+} + pyridin + oxygen of air, or from a solution of Fe^{2+} + hexamethylentetramine + $Na\,NO_2$.

From the above it seems that γFe_2O_3 is not a common specimen in nature. However, it was proved by Dr. C. F. Weenig (Laboratory for Geology of the Agricultural University, Wageningen) that the small brown (limonite-like) concretions which were found by Moerman (1947) in the heath on the Veluwe, in localities with prehistoric remnants (potsherds, etc.), are γFe_2O_3. Furthermore, when analyzing town refuse for plant nutrients and clay minerals, which material consists of street and household refuse (including some from stoves and furnaces), γFe_2O_3 was found also.

Lastly γFe_2O_3 was detected in the so called "rodoorn" soils of the provinces of Groningen and Drenthe.[1] These are brown-red colored humic soils (clays, sands) with high, or with relatively high amounts of iron (see table 2). Sometimes vast iron hardpans (see table 2) consisting of lepidocrocite (γFeOOH)[2] concretions occur here in the subsoil which are explored for several purposes (adsorbent in gas plants, etc.).

[1] The author is indebted to J. Godijk (Technical Leader of the field experiments at this institute) for his valuable samples of "rodoorn" soils and iron concretions rich in humus.

[2] Iron hardpans in peats, which are formed by oxidation of ferrous-organic compounds by oxygen of air, consist of lepidocrocite = γFeOOH (see table 1 and plt. 2 fig. 8). This is in contrast with common iron hardpans which are mostly formed by oxidation

From the above it may be concluded that γFe_2O_3 is a common mineral in sediments rich in iron and organic matter, especially when this material is heated by fires, as in prehistoric localities, household refuse, and peats which were set on fire to improve them for better cultivation of crops.

Details are given about the synthesis of γFe_2O_3 from material rich in iron and organic material; also, of γFe_2O_3 obtained synthetically by heating ferric citrate containing sodium as an impurity.

EXPERIMENTS[3]

As is mentioned above, γFe_2O_3 (see plt. 2, fig. 12 b and 12 c) was found in material rich in iron and organic matter, especially when this material was heated.

This phenomenon was a reason to try the synthesis of γFe_2O_3 in another way as was previously done by former investigators who only used inorganic components.

1.—Humic iron concretions rich in organic matter (see table 2), consisting of lepidocrocite (see table 1) from the "rodoorn" soils, were heated at ca. 600° C. for 2 hours. They changed in color from brown to red-brown and also became magnetic. (See plt. 2, figs. 9 and 12 d.) It was found by X-ray analysis (see table 1) that in this way a cubic substance was formed which was like γFe_2O_3 obtained from synthetic magnetite. (See plt. 1, fig. 2 and table 1.)

An X-ray pattern of γFe_2O_3 resembles that of magnetite obtained at low temperature (see plt. 1, figs. 2, 3 and 4 and table 1); also that of magnetite synthesized at high temperature by reduction of hematite with hydrogen, or that of magnetite of magmatic origin as found in nature. (See plt. 1, figs. 4 and 5 and table 1.)

The reason for this similarity is that the number of the vacant positions in γFe_2O_3[4] and in the low temperature magnetite[5] are comparatively so small that they cannot give a perceptible change in most of the intensities of the reflections, as compared with those of the high temperature magnetite.[6] Furthermore, the radii of OH^{1-} (1.33 A) and O^{2-} (1.32 A) of respectively the synthetic low temperature magnetite at the one hand and at the other hand the high temperature magnetite, and γFe_2O_3 are so small, that slight differences in the amount of these ions are also of little effect in most of the observed reflections.

However, γFe_2O_3 can be distinguished from magnetite obtained at low as well as high temperature, by means of a chemical analysis. The former does not contain ferrous iron and the latter has one-third part of all its iron ions in the ferrous form. (Theoretical = 31.03 per cent FeO.)

It was found by chemical analysis (see table 3) that the red-brown magnetic mineral found in the "rodoorn" soils, in prehistoric localities, as well as that obtained by heating lepidocrocite rich in organic matter at ca. 600° C., contained only ferric iron. Thus all three of these minerals are γFe_2O_3.

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of ferro(bi)carbonate by oxygen of air and which consist of limonite (goethite) = αFeOOH. (See table 1 and plt. 2 fig. 7.) Iron hardpans consisting of lepidocrocite, as well as of limonite are also found in nature.

[3] The author is indebted to Dr. E. W. Gorter (Philips Research Laboratories, Eindhoven) for his constructive critical remarks and his samples of minerals. The author is also indebted to Dr. P. Kruizinga (Laboratory for Geology and Mineralogy, Technical University of Delft) for his samples of goethite, limonite, lepidocrocite, hematite and magnetite, as pure as available in nature, and to J. D. Moerman (Apeldoorn) for his samples of soils with γFe_2O_3 and prehistoric remnants from the heath on the Veluwe.

[4] $\gamma Fe_2O_3 = 8Fe^{3+} - 1\frac{1}{3}Fe^{3+} - 32\ O^{2-}$; see Hägg (1935), Verwey (1935 a, b; 1936) and Kordes (1935).

[5] Precipitated magnetite (Fe_3O_4) is assumed by Starke (1939) to be $8Fe^{3+} - 8Fe^{3+}$, $5\frac{1}{3}Fe^{3+} - 26\frac{2}{3}O^{2-}$, $5\frac{1}{3}OH^{-}$.

[6] High temperature magnetite (Fe_3O_4) of magmatic origin or synthesized from hematite by reduction with hydrogen = $8Fe^{3+} - 8Fe^{3+}$, $8Fe^{2+} - 32\ O^{2-}$; see Hägg (1935) and Verwey (1936).

TABLE 1.—*X-ray diffraction data and indices of cubic minerals: γFe_2O_3 (synthetic, natural and ferric-citrate), magnetite = Fe_3O_4 (synthetic and natural). Furthermore, X-ray diffraction data of: hematite = αFe_2O_3, lepidocrocite = $\gamma FeOOH$ (pure and iron concretions), and goethite = limonite = $\alpha FeOOH$.* (Data summarized in decreasing order of their d spacings in A.)

γFe_2O_3 synthetic				γFe_2O_3				Fe_3O_4 (magnetite)				In-dices (hkl)[2]	Hematite αFe_2O_3		Lepidocrocite $\gamma FeOOH$				Goethite Limonite $\alpha FeOOH$	
magnetite *heated*		concretions *heated*		natural *soil*		ferric-citrate *heated*		synthetic (x)		natural					pure		iron *concretions*			
i	d	i	d	i	d	i	d	i	d	i	d		i	d	i	d	i	d	i	d
								VI	*11.5	VI*	11.5				VI	10.7				
										VI	7.59	1			VI	7.63			VI	(11.0
															IV	(6.57	IV	(6.57		
V	6.19	VI	6.19	VI	6.19			VI	6.23	VI	6.23	2								
										VI	*5.72									
															VI	5.12	VI	5.12		
V	4.97	VI	4.97	VI	4.97	IV/V	5.02	IV/V	4.99	IV	4.99	3							VI	5.02
										VI	*4.73									
										VI	*4.53									
VI	4.27	VI	4.27	VI	4.27	VI	4.33			VI	4.28	4			IV	4.28	IV	4.28	I	4.22
															VI	*3.91*	VI	*3.91*		
V	3.75	VI	3.75	V	3.75	VI	3.78			VI	3.76	5	III	3.70						
V	3.41	VI	3.41	V	3.41	VI	3.42			VI	3.41	6							V	3.40
															IV	3.38	IV	3.38		
VI	*3.25	VI	*3.25	VI	*3.25					VI	*3.25									
III	2.98	III	2.98	III/IV	2.98	III	3.01	III	3.00	III	3.00	8			VI	3.00	VI	3.00		
VI	2.82					VI	2.84			VI	2.83	9								
		VI	*2.70	V	*2.70								I	2.70	V	2.72	V	2.72	III	2.70
VI	2.66					VI	2.69			VI	2.68	10								
I	(2.54	II	(2.54	II	(2.54	I	(2.55	II/I	(2.55	II	2.55	11	II	2.51					V	2.58
VI	2.43	VI	2.43	VI	2.43	VI	2.46	VI	2.45	VI	2.45	12			III	(2.48	III	(2.48	II	2.45
VI	2.33	VI	2.33	VI	2.33	VI	2.37					13			VI	2.39	VI	2.39		
VI	2.23	VI	2.23	VI	2.23	VI	2.25	VI	2.24	VI	2.24	14			VI	2.28	VI	2.28	IV	2.25
										VI	*2.21		III	2.20	VI	2.20	VI	2.20	IV	2.19
III	2.09	III	2.09	IV/V	2.09	III	2.12	III	2.10	III	2.10	16	VI	2.10	VI	2.12				
								VI	2.04	VI	2.04	17			VI	2.08				
										VI	*2.02									
						VI	2.00					18							VI	2.00
				VI	*1.99															
		VI	1.90					VI	1.94	VI	1.94	19			IV	1.94	V	1.94	V	1.92
						VI	1.89	VI	1.88	VI	1.88	20			VI	1.85				
V	1.83	VI	1.83	VI	1.83					VI	1.83	21	III	1.84	VI	1.80	VI	1.80	V	1.80
										VI	*1.77									
IV	1.70	V	1.70	IV	1.70	IV	1.73	IV	1.72	IV	1.72	24			IV	1.73	IV	(1.73	III	1.71
						VI	1.69					25	II	1.69	VI	1.69			VI	1.68
						VI	1.66	VI	1.66	VI	1.66	26							VI	(1.65
III	1.61	VI	1.61	IV	1.61	III	1.63	III	1.61	III	1.61	27	IV	1.60	VI	1.62			VI	1.60
						VI	1.59					28								
VI	1.57			VI	1.57					VI	1.57	29			VI	1.57	VI	1.57	IV	1.56
		V	(1.54																	
VI	1.52			VI	1.52	VI	1.55			VI	1.52	30			IV	1.52	V	(1.52	IV	1.51
II	1.48	III	(1.48	III	(1.48	II	1.49	III	(1.48	III	1.48	32	III	1.49	VI	1.49				
				VI	*1.45															
				VI	1.42	VI	1.42	VI	1.45	VI	1.45	34	III	1.45	VI	1.46	VI	(1.46	IV	1.46
								VI	1.42	VI	1.42	35			VI	1.42			VI	1.42
VI	1.38	VI	1.38	VI	1.38	VI	1.38	VI	1.38	VI	1.38	37			VI	1.39			VI	1.39
				VI	*1.35			VI	*1.35	VI	*1.35		VI	1.35	VI	1.37			VI	(1.36
V	1.32	V	1.32	VI	1.32	V	1.33					40							V	1.32
								V	1.32	IV	*1.32	41								
						VI	1.31	VI	1.30	VI	1.30	42	IV	1.31	VI	1.31				
															VI	1.29			VI	1.29
IV	1.27	V	1.27	VI	1.27	IV	1.28	IV	1.27	IV	1.27	43			VI	1.27	VI	(1.27		
													V	1.26					VI	1.26
VI	1.25	VI	1.25	VI	1.25	VI	1.27	VI	1.26	VI	1.26	44								
										VI	1.25	45								
				VI	1.23	VI	1.24	VI	1.23	VI	1.23	46								
													VI	1.23	VI	1.24	VI	(1.24	VI	1.24
						V	1.21					48								
V	1.20	VI	1.20	VI	1.20			V	1.21	V	1.21	49	VI	1.22	VI	1.20	VI	(1.20	VI	1.20
				VI	1.18					VI	1.19	50	V	1.19						
VI	1.17	VI	1.17	VI	1.17	VI	1.18					51	V	1.17	VI	1.18	VI	(1.18		
								VI	1.17	VI	1.17	52								
VI	1.14	VI	1.14	VI	1.14	VI	1.15	VI	1.15	VI	1.15	54	IV	1.14	VI	1.14			VI	(1.14
						IV	1.12					56			VI	1.12			V	1.12
V	1.11	VI	1.11	VI	1.11			V	1.12	V	1.12	57								
						VI	1.11	VI	1.10	VI	1.10	58	IV	1.10						
IV	1.08	V	1.08	VI	1.08	III	1.09	IV	1.09	IV	1.09	60			VI	1.09			VI	1.09
						VI	1.08	VI	1.08	VI	1.08	61	VI	1.08						
						VI	1.07					62			VI	1.07			VI	1.07
								VI	1.06	VI	1.06	64								
V	1.04	VI	1.04	VI	1.04	IV	1.05	V	1.05	V	1.05	65	IV	1.05	VI	1.05			VI	1.05

TABLE 1—(*continued*)

γFe_2O_3 synthetic				γFe_2O_3				Fe_3O_4 (magnetite)				Indices $(hkl)^2$	Hematite αFe_2O_3		Lepidocrocite $\gamma FeOOH$				Goethite Limonite $\alpha FeOOH$	
magnetite heated		concretions heated		natural soil		ferric-citrate heated		synthetic (x)		natural					pure		iron concretions			
i	d	i	d	i	d	i	d	i	d	i	d		i	d	i	d	i	d	i	d
VI	1.03	VI	1.03			VI	1.04					66	V	1.04	VI	1.04				
										VI	1.03	67			VI	1.02			VI	1.02
										VI	1.01	69			VI	1.01			VI	1.01
V	0.98	VI	0.98			V	0.99	V	0.99	V	0.99	73	V	0.99	V	0.99			VI	0.99
													VI	0.97						
VI	0.96	V	0.96	VI	0.96	IV	0.97	IV	0.97	IV	0.97	76	IV	0.96						
										V	0.96	77								
		VI	0.95	VI	0.95							78	V	0.95	VI	0.96				

i=relative intensities estimated visually in grades: I=very strong, II=strong, III=medium, IV=weak, V=very weak, VI=barely visible.
d=interatomic spacing. *=impurity. (=reflection broadened (band).
(x) low temperature magnetite with bands; high temperature magnetite without bands.

2.—Lepidocrocite or goethite or limonite (limonite found in nature=$\alpha FeOOH$, is said to be goethite=$\alpha FeOOH$ with excess of water), free from organic matter when heated at ca. 600° C., are converted to hematite (see plt. 2, fig. 6 and table 1). If, however, fine powdered lepidocrocite or limonite are thoroughly mixed with fine organic matter such as amylum, humic acid, etc., and afterwards heated for some hours at ca. 600° C., γFe_2O_3 is obtained. (See plt. 2, fig. 10.) Furthermore, it was found that γFe_2O_3 could also be produced if limonite concretions, free from organic matter, even of some millimeters in size, are heated together with citric or oxalic acid at ca. 600° C. for 2 hours. (See plt. 2,

TABLE 2.—*Chemical composition in per cent of iron hardpans and red colored soils from different localities in which γFe_2O_3 was found in natural state*

Constituents (per cent)	Iron Hardpans		Red colored soils		
	Zuidbroek (288)	Marum (657)	Meden (285)	Marum (286)	Ommen (222)
loss on ignition	33.6	31.6	29.5	49.4	13.8
(carbon)	(6.3)	(9.0)	(5.6)	(16.1)	(4.3)
SiO_2	4.1	15.6	16.6	22.9	62.5
Fe_2O_3	54.06	45.71	49.09	21.07	17.89
Al_2O_3	0.45	1.66	1.01	2.33	1.95
TiO_2	tr.	tr.	0.14	0.06	0.09
Mn_3O_4	0.05	tr.	0.06	tr.	tr.
P_2O_5	4.04	2.59	1.26	1.06	0.57
CaO	2.10	1.47	0.76	1.46	0.81
MgO	0.42	0.24	0.61	0.40	0.42
K_2O	tr.	tr.	0.42	0.20	0.45
Na_2O	0.30	0.11	0.28	0.18	0.36
Total	99.12	98.98	99.73	99.06	98.84
Other	0.88	1.02	0.26	0.94	1.16

tr.—traces=<0.05 per cent.

TABLE 3.—*Loss on ignition and Fe_2O_3 in per cent for: (a) natural γFe_2O_3 from soils from different localities; (b) synthetic γFe_2O_3 obtained by heating iron concretions rich in organic matter at ca. 600° C.; (c) natural magnetite from different localities*

No.	Locality	Loss on ignition (per cent)	Fe_2O_3 (per cent)	Total (per cent)	Other elements (per cent)
	(a) *natural γFe_2O_3* found in soils				
2850	Meden	10.5	87.3	97.8	2.2
6500	Marum	10.6	88.0	98.6	1.4
2200	Ommen	9.1	90.2	99.3	0.7
6350	Nw. Milligen	9.2	86.9	96.1	3.9
5870	Otterlo	7.1	87.3	94.4	5.6[1]
	(b) *synthetic γFe_2O_3* (heated iron concretions rich in organic matter)				
607	Marum	1.4	86.3	87.7	12.3[2]
	(c) *natural magnetite Fe_3O_4* (magmatic origin)				
611	Germany	1.4	90.1[3]	91.5	8.5[4]
142	Germany	traces	97.7[3]	97.7	2.3[5]

[1] CaO+MgO=1.68, P_2O_5=3.29, Al_2O_3=tr., Mn_3O_4=0.37, TiO_2=0.10.
[2] SiO_2=3.00, CaO+MgO=1.31, P_2O_5=5.03, Al_2O_3=2.42.
[3] Fe_2O_3 as Fe_3O_4.
[4] CaO+MgO=4.66, P_2O_5=3.52, Al_2O_3=tr., Mn_3O_4=0.15, TiO_2=tr. (apatite-rich magnetite).
[5] CaO+MgO=1.12, P_2O_5=tr., Al_2O_3=0.62, Mn_3O_4=0.05, TiO_2=0.40 per cent.
tr.=traces=<0.05.

fig. 12 e.) From the above it may be concluded that, to get γFe_2O_3, the reducing organic matter should be in narrow contact (or that reducing organic fumes could make narrow contact) with lepidocrocite or limonite, as otherwise both are converted to hematite.

3.—Glittering lamellae of ferric citrate ($C_6H_5O_7Fe \cdot 3H_2O$), of the quality as commonly sold by druggists, were heated for 2 hours at ca. 600° C. They changed to dull red-brown magnetic lamellae. (See plt. 2, fig. 12 f.) It was found by X-ray analysis (see plt. 2, fig. 11 and table 1) that a cubic substance was formed, which apart from slight differences in the intensities of some reflections and the size of the unit cell, was similar to γFe_2O_3. By chemical analysis it was further proved that only traces of ferrous iron are found. However, this substance contained 5.26 per cent Na_2O of which 0.35 per cent Na_2O and 0.86 per cent Na_2O were, respectively, in the form of NaCl and Na_2SO_4.

DISCUSSION

According to Hägg (1935) three possibilities may occur for the structure of γFe_2O_3 in the close-packed oxygen lattice:

(a) vacant Fe^{3+} positions in tetrahedral holes and intensity of (111) spacing=26 as the result;

EXPLANATION OF PLATE 1

FIG. 1.—Scale of X-ray photos (cm and A).
FIG. 2.—X-ray photo of synthetic γFe_2O_3 obtained by heating the synthetic low temperature magnetite of Lefort (1852).
FIG. 3.—X-ray photo of γFe_2O_3 found in nature.
FIG. 4.—X-ray photo of synthetic magnetite (Fe_3O_4) (low and high temperature).
FIG. 5.—X-ray photo of magnetite (Fe_3O_4) found in nature.
FIG. 6.—X-ray photo of hematite (αFe_2O_3) synthetic and as found in nature.
FIG. 7.—X-ray photo of limonite ($\alpha FeOOH$) concretions as found on the heath of The Veluwe.

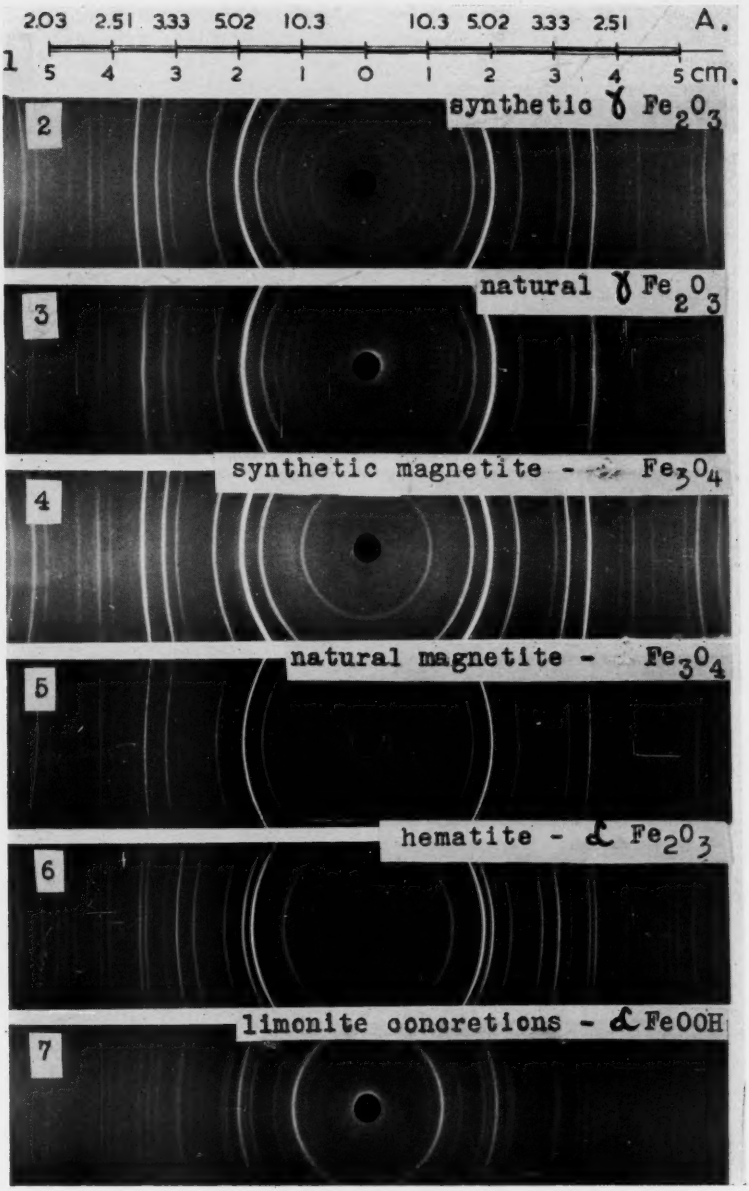
2.03 2.51 3.33 5.02 10.3 10.3 5.02 3.33 2.51 A.
1 5 4 3 2 1 0 1 2 3 4 5 cm.
2
synthetic γ Fe_2O_3
3
natural γ Fe_2O_3
synthetic magnetite - Fe_3O_4
4
natural magnetite - Fe_3O_4
5
hematite - α Fe_2O_3
6
limonite concretions - α FeOOH
7

PLATE 1

(b) vacant Fe^{3+} positions in octahedral holes and intensity of (111) spacing = 1.4 as the result;

(c) vacant Fe^{3+} positions in the tetrahedral holes as well as in the octahedral holes and intensity of (111) spacing = 6.1 as the result.

Gamma Fe_2O_3 obtained by heating of synthetic magnetite at 200°–300° C. should be b (vacant Fe^{3+} places in octahedral holes) as was found by Verwey (1936) by means of intensity calculations. This mineral is found also in nature in the Dutch "rodoorn" soils, or it may be obtained by heating concretions of lepidocrocite or limonite rich in humus to ca. 600° C. By the latter procedure the slow process of synthesis as happens in nature, is hastened. Accordingly, more γFe_2O_3 particles are found in material rich in organic matter and limonite, or lepidocrocite, when these are heated by fire; for example in household refuse, in prehistoric localities, or where iron-rich peats are burned to improve them for better cultivation of crops.

A peculiarity of this γFe_2O_3 is that it is more stable against heating as γFe_2O_3 from synthetic magnetite, as the latter one is easily converted to αFe_2O_3 when heated to 400° C.

We know, however, from the experiments of Michel and Chaudron (1935) and those of Bénard and Chaudron (1938), that even small amounts of sodium and manganese as occur in the lepidocrocite and limonite concretions (see tables 2 and 3), may stabilize γFe_2O_3. The presence of sodium was also the reason that γFe_2O_3 obtained from the impure ferric citrate could withstand heating for 2 hours by 600° C.

The unit cell of magnetite and γFe_2O_3 calculated from the X-ray reflexions is[7]

a_0 Fe_3O_4 (synthetic and natural) = 8.45 A[8]
$a_0\gamma Fe_2O_3$ (synthetic and natural) = 8.40 A[9]
$a_0\gamma Fe_2O_3$ (heated ferric citrate) = 8.48 A

As may be noted by comparison of the above figures with those found by other authors (see footnotes 8 and 9) differences occur. The reason for these differences may be ascribed justly to the kind and the amount of impurities investigated in the samples. It is known from the experiments of Michel and

[7] According to the Statistical Department of the Central Organization for Applied Research in the Netherlands T.N.O., The Hague, it was found by means of a least square calculation from the observed reflections which were assumed to have the same weight: $a_0Fe_3O_4 = 8.441 \pm 0.0043$ A; $a_0\gamma Fe_2O_3 = 8.393 \pm 0.0044$ A and $a_0\gamma Fe_2O_3 = 8.476 \pm 0.0050$ A.

[8] According to Hägg (1935), = 8.380 A; to Michel (1937), = 8.41 A; to Bénard (1939), = 8.38 A; to Frank (1939), = 8.42 A; and to Verwey, = 8.40 A (1936) and = 8.39 A (1947).

[9] According to Hägg (1935), = 8.332 A; to Verwey (1935a), = 8.40 A; to Michel and Chaudron (1935), = 8.32 A; and to Haul (1939), = 8.32 A.

EXPLANATION OF PLATE 2

FIG. 8.—X-ray photo of lepidocrocite (γFeOOH) concretions rich in organic matter as found in the brown-red colored ("rodoorn") soils of the provinces of Groningen and Drenthe.

FIG. 9.—X-ray photo of lepidocrocite concretions rich in organic matter heated at ca. 600°. C for 2 hours = γFe_2O_3.

FIG. 10.—X-ray photo of pure powdered limonite (αFeOOH) heated with amylum at ca. 600° C. for 2 hours = γFe_2O_3.

FIG. 11.—X-ray photo of lamellae of ferric citrate heated at ca. 600° C. for 2 hours = γFe_2O_3.

FIG. 12a.—γFe_2O_3 obtained by gentle heating (200°–300° C.) of synthetic magnetite according to the method of Lefort (1852).

b.—γFe_2O_3 found in nature at localities with prehistoric remnants on the heath of The Veluwe.

c.—γFe_2O_3 found in nature in the brown-red tinted soils ("rodoorn" soils) of the provinces of Groningen and Drenthe.

d.—γFe_2O_3 obtained by heating lepidocrocite concretions rich in organic matter at ca. 600° C. for 2 hours.

e.—γFe_2O_3 obtained by heating limonite concretions (free from organic matter) with oxalic acid at ca. 600° C. for 2 hours.

f.—γFe_2O_3 obtained by heating lamellae of ferric citrate at ca. 600° C. for 2 hours.

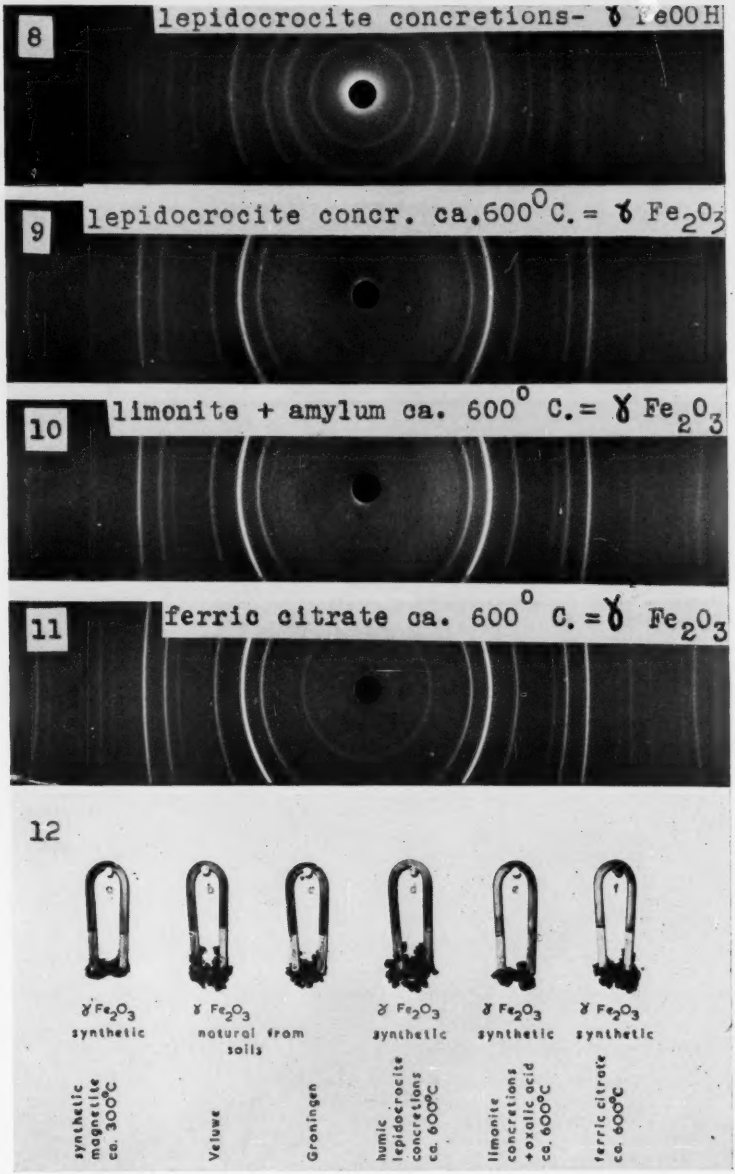
8
lepidocrocite concretions- γ FeOOH
9
lepidocrocite concr. ca.600°C. = γ Fe_2O_3
10
limonite + amylum ca. 600° C. = γ Fe_2O_3
11
ferric citrate ca. 600° C. = γ Fe_2O_3
12
a
b
c
d
e
f
γ Fe_2O_3 synthetic
synthetic magnetite ca. 300°C
γ Fe_2O_3 natural from soils
Veluwe
Groningen
γ Fe_2O_3 synthetic
humic lepidocrocite concretions ca. 600°C
γ Fe_2O_3 synthetic
limonite concretions + oxalic acid ca. 600°C
γ Fe_2O_3 synthetic
ferric citrate ca. 600°C

PLATE 2

Chaudron (1935, 1949) that even small amounts of sodium are cause of a larger unit cell of γFe_2O_3. This is due to the forming of a mixed structure of the type $(Fe_{1-x}Na_x)_2O_3$.

Sodium is also the reason that the unit cell of γFe_2O_3 obtained from ferric citrate, which contained in total 5.26 per cent Na_2O (in this total is included 0.35 per cent Na_2O from NaCl and 0.86 per cent Na_2O from Na_2SO_4), was an even 8.48 A; this is very much larger than that of pure γFe_2O_3[10]. In contrast with this, small amounts of Al may diminish the unit cell of γFe_2O_3, as is shown by Michel and Pouillard (1949).

The unit cell of magnetite is augmented by small amounts of titanium and manganese by which the type $Fe_2O_3[(FeO)_{1-x}|(Ti^{or}\ Mn)O|_x]$ is formed, as was proved by the experiments of, respectively, Pouillard (1949) and Bénard and Chaudron (1937). The results of the French investigators who worked on the subject of mixed ferrites and γFe_2O_3 are recently summarized by Michel *et al.* (1950).

Fe_3O_4 (magnetite).—This substance, when synthesized from solutions at low temperature, is easily (ca. 200°–300° C.) oxidized in the air to γFe_2O_3. During this oxidation some contraction of the unit cell takes place as a result of the transformation of the larger Fe^{2+} ions (0.78 A) to the smaller Fe^{3+} ions (0.67 A) and the even smaller holes. At higher temperatures (ca. 600° C.), magnetite obtained by synthesis at low temperature, is converted first to γFe_2O_3 and thereafter to hematite. As is mentioned by Krause (1931) and Fricke (1937), respectively, the amount of Fe^{2+} and the size of the particles of the low temperature magnetite, influence its stability against heating (oxidation). According to Baudisch and Welo (1926), differences in chemical and physical (among them magnetism) properties of the low temperature magnetites obtained in several ways, are caused by irregularities in the crystal lattices.

Magnetite from magmatic (volcanic) origin, however, or magnetite obtained synthetically by reduction of hematite at high temperature (ca. 1400° C.) with hydrogen, is a very stable product. According to Welo and Baudisch (1925), this mineral is not converted to hematite until at ca. 800° C. This might be explained by the assumption made by Starke (1939) concerning the structure of precipitated magnetite. (See footnote 5.)

[10] The sample of γFe_2O_3 which was obtained from synthetic magnetite according to the method of Lefort (1852), still contained 0.09 per cent Na_2O. Gamma Fe_2O_3 having a unit cell of 8.52 A and containing 5 per cent Na_2O was obtained by Michel and Chaudron (1935).

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FIGS. 1–3, MARCH, 1951

SEA LIONS AS GEOLOGICAL AGENTS

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ABSTRACT

Sea mammals, although known to carry gastrolith pebbles in their stomachs, have not hitherto been considered important geological agents. Numbers of water worn pebbles on the surface and embedded in Recent peat deposits at the Snares Islands south of New Zealand, are attributed to transport by sea lions. Erratic pebbles of basalt have apparently been carried 130 miles to the Snares Islands from the Auckland Islands.

K. O. Emery (1941) has emphasized "the efficacy of sea mammals in distributing pebbles carried as stomach stones," described pebbles from the stomachs of California sea lions, and concluded that sea mammals "cannot be quantitatively important as geological agents" although they may account for "infrequent pebbles in fine-grained sediments." During geological examination of the Snares Islands, south of New Zealand, the writer found evidence that sea lions may locally be important geological agents in transporting "erratics" long distances from their source and in depositing marine pebbles in terrestrial beds where they may accumulate in large numbers.

The Snares are a group of subantarctic islets lying 62 miles S.S.W. of Stewart Island, the southernmost and least of the three main islands of New Zealand (fig. 1). The main island, some 2 miles in length, 620 ft. in height, is a cliff-bounded mass of Paleozoic granite, truncated by well-preserved remnants of a low-relief erosion-surface of late-Tertiary or post-Tertiary age (fig. 2), modified by stream erosion and by wave attack.

The islands lie in latitude 48° S and did not suffer glaciation during the Pleistocene. Precipitation is high and the islands support a rich vegetation of subantarctic tussock and coastal forest dominated by the tree daisy, *Olearia lyalii*, growing on a discontinuous soil of water-logged humus ("peat") post-glacial in age, locally exceeding 20 ft. in depth. Pollen analysis by W. M. Harris indicates a relative decrease in forest and increase in tussock during deposition of the uppermost 13 feet of peat, which may thus be no older than the post glacial climatic optimum, several thousand years B.C.

Smooth water-worn pebbles, 1–4 inches in length, lie scattered diffusely on the surface of the peat within about one-half mile of landing places on the east coast, at heights up to about 200 ft. above sea level. Natural sections rarely expose isolated rounded pebbles lying deeply embedded in the peat. Colonial nesting penguins (*Eudyptes pachyrhynchus atratus* Finsch) have destroyed protecting vegetation and by constant tramping have helped rain-wash to remove the peat cover and locally bare the rock surface. In such places rounded pebbles have accumulated as a lag gravel.

Most of the water worn pebbles are composed of local granite, but a small proportion (judged at less than one in a hundred) are of fine-grained dark basalt. The Auckland Islands, 130 miles S.S.W. of the Snares and separated by water exceeding 100 fathoms in depth, are the nearest known source of basalt. The Snares pebbles, lying diffusely on and within a layer of post-glacial peat, cannot be attributed to Pleistocene shore drift

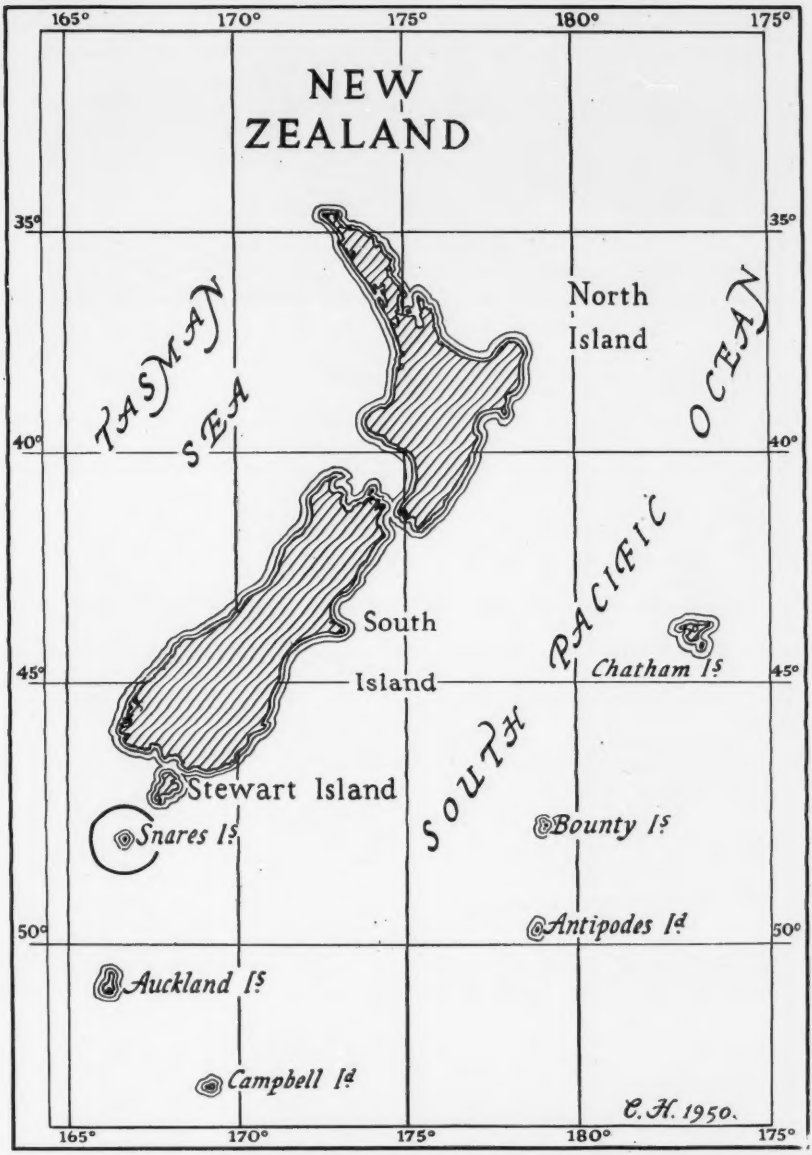

FIG. 1.—Map of southern New Zealand and outlying islands, to show the position of the Snares and Auckland Islands.

FIG. 2.—Northern promontory of the main Snares Island, showing remnant of summit plateau (skyline), cliffed west coast (left) and vegetal cover of tussock and scrub growing on thick peat. Sea lion gastroliths occur in peat exposed in the saddle (x) and in penguin colonies accessible from the east coast. Drawn from a photograph.

or ice transport.

Hooker's sea lion (fig. 3), *Zalophus hookeri* (Gray), has its chief breeding place at the Auckland Islands, and the animals, chiefly immature, which frequent other New Zealand subantarctic islands, the coasts of Stewart Island, and Foveaux Strait, are classified as a non-breeding overflow from the Auckland Island population. It is not known how regularly nor how rapidly sea lions make the passage between the Auckland Islands and the Snares, but it may be safely inferred that they do so.

Sea lions habitually take up and swallow pebbles, probably, as suggested by Turner (1887), to assist the trituration of food in the stomach; alternative suggestions have been listed by Hamilton (1934) and Emery (1941). Probably they obtain the pebbles on the shallow sea bottom while diving and groping for their animal food; they have not been observed to take them up on land. After a period of feeding at sea, the animals come to land and lumber up the slopes into coastal scrub and forest, sometimes to a height of more than 600 ft. (Auckland Islands). On land, sea lions lie sleeping and drowsing for several days, and during this interval, the gastrolith pebbles are ejected, commonly with undigestible food remains (cephalopod, gastropod, crustacean, fish). The ejection of gastroliths is accompanied by regurgitatory movements, belching and grunts, which culminate in the projection of a group of pebbles, several pounds in total weight. Sometimes the pebbles are deposited as a compact group, but violent shaking of the head during the process may splatter them over an area of several square yards, where they lie on the surface of the peat and may be pressed into it by the subsequent movement of the animals across them.

The above account of the behaviour of Hooker's sea lion is based on personal observation at the Auckland and Snares Islands where the habit of ejecting stones on land has been frequently observed. Stones and gravel are commonly recorded from the stomachs of other Pinnipedia (Turner, 1887; Wilson, 1907; Hamilton, 1934; Bertram, 1940; other references quoted by Emery, 1941). The ejection of gastroliths on land has seldom been described. Indeed, Hamilton (1934) stated that he had never observed the Falkland sea lions throw up its stomach pebbles.

At the Snares, no freshly ejected pebbles longer than about 2½ inches were seen and no freshly ejected basalt pebbles were found. Emery (1941), quotes a record of "single rocks weighing as much as five pounds" from the stomach of an unspecified seal, and Hamilton

FIG. 3.—Hooker's sea lion, *Zalophus hookeri* (Gray), adult male, photographed during a period of resting on shore at Rose Island, Auckland Islands, September 1942.

measured one stone 94 mm (3¾ inches) long, from the stomach of the Falkland sea lion (*Zalophus byronia*). It seems probable that the largest rounded pebbles at the Snares could have been ejected by a sea lion, and that the basalt pebbles have been carried from the Auckland Islands as gastroliths.

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FIGS. 1–2, MARCH, 1951

AN UNUSUAL STALACTITE FROM SALTVILLE, VIRGINIA

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ABSTRACT

The stalactite described and illustrated shows an unusual development of a single crystal of calcite expressed in external as well as internal form.

A stalactite with well-developed hexagonal prismatic crystal form was recently brought to the writer by T. M. N. Lewis professor of anthropology at the University of Tennessee. The stalactite was sent to Mr. Lewis by its discoverer, Mr. Rufus W. Pickle of Saltville, Virginia. It was found in a cave about 15 miles from Saltville.

The upper part of the stalactite bears the radial structure and irregular form found on most stalactites, but the lower part, representing nearly one half the total length, is a single crystal unit (fig. 1). The crystal has the form of a slender prism tapering downward from a cross section diameter of 10 cm to 7 cm. The crystal was broken when received so that the perfect $10\bar{1}1$ cleavage of calcite was revealed. The major crystal form is the prism $(11\bar{2}0)$; its faces are well-developed, nearly equal in size, and bear growth striations parallel to the trace of the $(10\bar{1}1)$ plane. The crystal also shows a very slight development of the $(10\bar{1}0)$ form.

Viewed in cross section the crystal exhibits an outer zone or shell about 1 mm thick composed of white translucent calcite. Clear glassy calcite fills this shell except for the central vent which runs the length of the crystal. Although the central portion and the shell are a part of the same crystal unit, they differ so markedly in appearance as to suggest that they were not deposited contemporaneously.

Stalactites showing the cleavage of a single crystal unit are not uncommon, but those with the external form of a crystal are apparently rare. The difference between the inner and outer parts of the crystal (fig. 2) leads to the suggestion that the inner zone formed first as a single crystal unit but with irregular external form. Calcite deposited later on the outer part of the crystal assumed a parallel crystal structure resulting in an euhedral form. Although the two stages were apparently distinct, one may have followed immediately after the other so that the development of the hexagonal form nearly kept pace with growth of the core.

FIG. 1.—Stalactite with hexagonal prismatic crystal form.

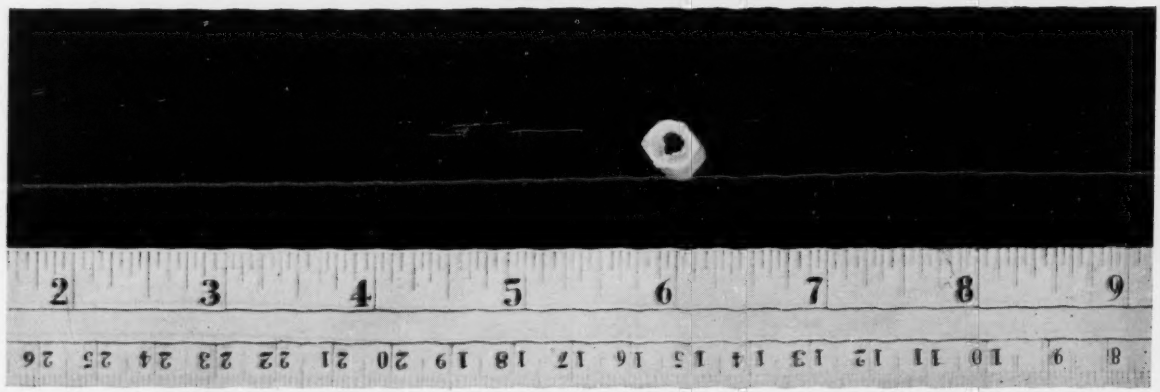

FIG. 2.—Cross section of stalactite. Upper surface is cleavage plane revealed as fracture on fig. 1.

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MARCH, 1951

THE PROBLEM OF DOLOMITE FORMATION CONSIDERED IN THE LIGHT OF RESEARCH ON DOLOMITES IN THE SONNWENDMOUNTAINS (TIROL)

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ABSTRACT

The processes of dolomite formation and the time of their occurrence in the history of a sediment are discussed. Pertinent evidence is presented from the writer's studies in the Sonnwendmountains and from other sources.

The following dolomites have been demonstrated by B. Sander (1936, p. 177): 1) mechanical sedimentation, external and internal in cavities; 2) chemical sedimentation (growth), external and internal in cavities; 3) biogenic sedimentation (growth), external; 4) metasomatic processes in the sediment.

The present author has intentionally avoided the word "primary" in dolomite processes, as this might lead to confusion when discussing the question of whether or not we are dealing with the first origination of dolomite. If, for instance, we speak of a "primary" dolomite in the sense of its first formation as dolomite, that may equally well be an authigenic as an allothigenic formation of dolomite. A "primary" dolomite may be settled allothigenic dolomite sand or it may be the filling of a cavity. An original metasomatic dolomite formed as dolomite (but secondarily arisen from lime) is "primary" when considered as dolomite, but "secondary" when considered as replacement of the lime (after Sander). According to Cloud and Barnes (1948, pp. 45–50), as well as in the opinion of the present author, the following manner of dolomite formation may be identified in recent material: 1) thrown up dolomite settled on secondary deposition (or also called "dolomite-resediment"), external and internal in cavities, as may be found in certain coral islands in the Pacific and Indian Oceans; 2) chemical sedimentation, external and internal, resulting from the evaporation of an enclosed water system—such dolomite is usually accompanied with gypsum and several evaporite salts; 3) biogenic sedimentation, external.

It should be noted as an introduction that a siliceous solution in the sediment reacts differently upon calcareous organodetrital remains, than it does to the lime sediments in which these remains are embedded. The siliceous solution in the material of the Sonnwendmountains, for example, only effected a diffuse impregnation in a sediment which was in the earliest stage of consolidation. If the sediment had reached a later stage of consolidation, no diffuse impregnation took place. The siliceous solution was able, however, to diffusely impregnate solid calcareous organodetrital remnants and to completely silicify them while retaining the structure. In a thin section of Koessenerstrata (upper Triassic) a foraminifera of the Lagenidae was found, the shell consisting entirely of dolomite after a pseudomorphism from lime to dolomite. It can be seen plainly in the same thin section, also, how oolitic lime has been replaced by dolomite. In general these dolomite metasomatoses in the oolitic lime were very coarse grained, so that the radial lamellary structure of the oolitic lime is badly preserved and often disappeared after the metasomatic process. The fine structure of the organodetrital remnants, however, has often been preserved even after the pseudo-

morphism from lime to dolomite. Further research in this direction is certainly very much to be desired.

Maegdefrau (1942, p. 203) noted the important fact that in the calcareous algal beds in the Gulf of Naples, the thallus layers of the upper living calcareous algae show a very small content of Mg, and that the increase of Mg occurs only in the lower (dead) thallus layers, a fact that he ascertained by means of Lemberg stain. (If a biogenic fragment has such a small amount of Mg in it that it remains practically black after Lemberg stain, it is not here regarded as a dolomite.)

In the Indian and Pacific Oceans and, also, in the Soenda Sea, are still corallinacea and dasycladacea which show a small amount of dolomite, usually to 5 per cent, occasionally 10 per cent. In fossil material, on the other hand, we should be cautious in concluding from a dolomitic biogenic fragment that the organism contained more or less Mg when living. To be certain of this, an exact petrographic study, with the help of thin and polished sections, on sediments must be made, as experience has shown.

Maegdefrau found that living algae with a low Mg content, slightly dolomite-containing, remain practically black after the Lemberg stain. A somewhat further advanced stage of dolomitization, on the other hand, can be recognized immediately by the same method. *Thus where a dolomitic-biogenic fragment can be detected in thin sections by means of Lemberg stain, we may assume with some certainty that the increase of dolomite has taken place only after death.*

Brauch (1923, pp. 179–181) came to the conclusion that the dolomite of the Zechsteinreef (Germany) is not "primary," in so far that the algae and corals did not contain enough Mg when they were living to be called dolomites, but that the dolomitic character of the reef set in later; that is, immediately inceptive in the algae layers as soon as they were dead.

4) No examples of primary deposition of authigenic dolomite are known. The question arises where to find the large and massive dolomites of the most recent times comparable to the extensive dolomites of especially Triassic time. In the Sonnwendmountains it could be demonstrated that the great masses of dolomites were formed subaquatically during the process of parainduration[1] of the sediments. The composition of the Jurassic intergranular sea water, for instance, must have possessed a large solvability for Mg, as this solvability in not yet consolidated sediment is very high on existing allothigenic dolomite components. The allothigenic dolomite components are often broken in consequence of deformation which took place during the latest stages of consolidation of the sediment, or in other words, during the process of parainduration. The fissures are then filled up by dolomite which was provided by the allothigenic dolomite components themselves. Around the allothigenic dolomite components which lie in a calcareous sediment, the intergranular waters can dissolve a part of the dolomite content and can proceed metasomatically along the lines of least resistance in the sediment (belteropore advancement; Sander, 1936, p. 31, 47). This happens very often. On the other hand, the formation of dolomite rhombohedra is a distinctly authigenic process and signifies the origin of dolomite in the sediment. In the material of the Sonnwendmountains it took place only during parainduration. As soon as the process of parainduration is completed and thus, also, the rock, no further formation of dolomite rhombohedra took place.

Metasomatic activity in other forms may take place in all diagenetic stages (Weynschenk, 1949, chap. II). When the dolomite has a granular structure, we are certain that we are dealing with a settled allothigenic dolomite sand only if the dolomite grains are intergranularly bound together by calcite. If the intergranular

[1] Lithification of other authors.

space also consists of dolomite, it may have been a dolomite sand cemented intergranularly by dolomite, but, it may also be an original authigenic dolomite formation in the sediment. The two formations, at any rate in the present rich materials, cannot be distinguished. The coarseness of grain is positively not related to irregularity of dolomitization and its cause has to be investigated (see also Cloud and Barnes, 1948, p. 63).

The most satisfactory method of investigating dolomites is by the Lemberg hydroxide stain (Rodgers, 1940, p. 789). The specimen is dipped in $(NH_4)_2S_x$ and the stained surface is put under a fine jet of water and then covered with glycerine (which is diluted with one-third part water). The preparation remains good for more than three-quarters of an hour; as long as the glycerine-cover is on the stain the black FeS cannot oxidize and the disadvantages (described by Rodgers, 1940, p. 790) of unstable oxidizing FeS turning brown and crumbling and leaving cracks all over the calcareous areas, are overcome. It is also better to decolourize the preparation by washing with a 2 to 3 per cent HCl solution by means of a wad of cotton wool. Before and after the act of decolourizing, it may be necessary to put the preparation under a fine jet of water. By this method a specimen or thin section may be stained many times.

An increase of dolomite took place in the so-called "Oberjura-Vorfeld breccia"; that means in a sea region with a depth of about 10 m to 50 m maximum in front of the coastal "Hornsteinbreccia," which has mainly a depth of about 10 m and less. The "Hornsteinbreccia" is marked by mostly allothigenic metasomatic dolomitization processes of which the allothigenic dolomite grains represent the dolomite. Authigenic processes were seldom met and not intensive at all. Therefore, the allothigenic secondary dolomitization processes, as well as authigenic processes, in the "Oberjura-Vorfeldbreccia" underwent an enormous increase in certain regions, as in the "Latschboden." The author does not regard the thickness of the water layer as of influence for the dolomitization processes, according to his own investigations and the literature. (See Cloud and Barnes, 1948, p. 62.)

The subaquatic diagenesis, with its several stages of consolidation during parainduration, occupies a great amount of time, and the chemical processes in the sediment, due to the transportation of material through the intergranular system, proceeds very slowly, Therefore, in recent or quite young sediments only small amounts of dolomite can be found on the periphery. The best proof is the investigation on Funafuti atoll, where it is only at a depth below 200 m that enough dolomite was found to speak of dolomites. (See Andree, 1920, pp. 171–172.)

When Maegdefrau found some dolomites in the dead thallus layers of the Gulf of Naples, it was noted this particular dolomite process had needed less time than the dolomite formation in the sediment itself. Presumably the rapidity of the dolomitization process is effected by the difference in density and structure of the biogenic lime as compared to the calcareous composition and structure of the sediment. However, Dana's supposition in 1849 that dolomitization took place during parainduration, may certainly be confirmed. No authigenic dolomite has been formed on the free surfaces in the past and the Triassic dolomites were not formed in this way.

It is to be expected that dolomites may be found of Recent and young geological date of large mass and extent in the sediment of tropical seas from a few hundred meters depth towards greater depths. These will furnish the most recent comparative material to the extensive dolomites of ancient times.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 21, NO. 1, PP. 32–33
FIG. 1, MARCH, 1951

A COMPARISON CHART FOR VISUAL PERCENTAGE ESTIMATION

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It is often desirable to estimate mineral percentages in a rock, be it outcrop, hand specimen, or thin section. For accurate work, recourse may be had to a Rosiwal analysis or some other statistical type of measurement, but for rapid estimation the writer has found the accompanying chart (fig. 1) useful. Most workers have a tendency to overestimate small percentages, (particularly in the case of highly conspicuous minerals like mica), and comparison with this chart helps to reduce such errors. The writer has used it for the following purposes: obtaining percentages of minerals in hand specimens and thin sections of rocks; estimating proportions of rock types present in drill cuttings; determining the percentage of patches of calcite cement in sandstones by noting the effervescence; and estimating visible porosity in limestones and dolomites.

Each square contains a given percentage of black area, which percentage is printed below the square. Each quarter of the square contains the same amount of black, but the arrangement is varied. To use the chart, one may construct a paper mask with a hole cut in the center the same size as the squares; this is moved over the chart until a square is found in which the percentage of black area most nearly matches the percentage of the desired substance in the specimen.

This chart has been adopted as standard procedure in the Basic Sedimentology course (Mineralogy 511–512) taught at the Pennsylvania State College.

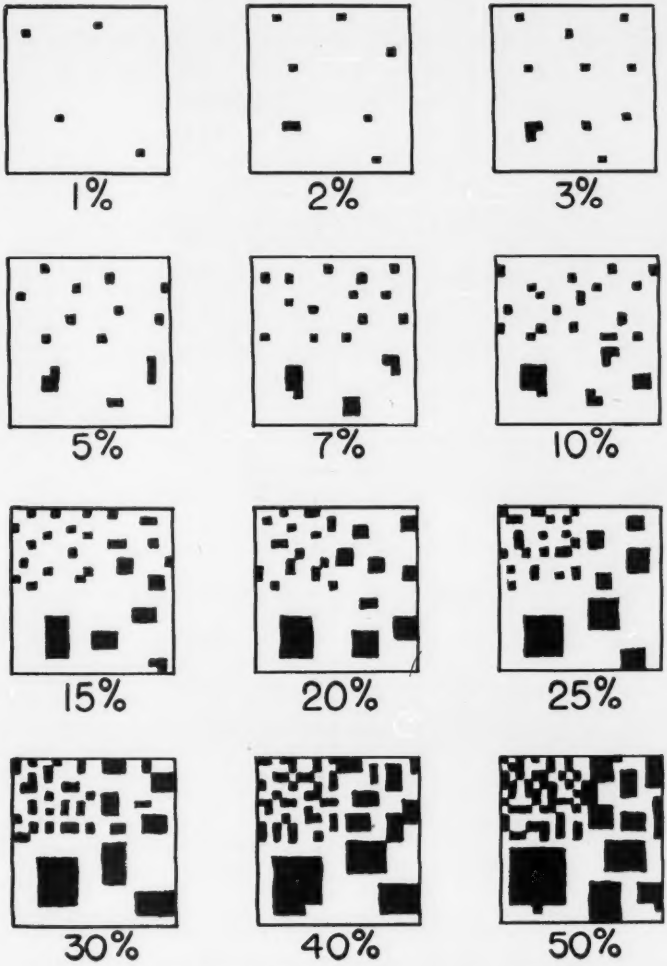

FIG. 1.—A comparison chart for visual percentage estimation.

JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 21, NO. 1, PP. 34–46
FIGS. 1–3, MARCH, 1951

SOME OCCURRENCES OF SUPERGENE IRON SULPHIDES IN RELATION TO THEIR ENVIRONMENTS OF DEPOSITION

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ABSTRACT

The widespread development of marcasite in association with the acid environment of brown coal deposits is in striking contrast to the development of pyrite in marine clays that provided a neutral or alkaline environment, and suggests that the form of ferrous sulphide found in supergene deposits is an indicator of the acidity or otherwise of the conditions attending its decomposition. The replacement of humified wood by marcasite, with perfect preservation of the wood structure, and the filling of marine shells by pyrite seems best accounted for by the action of sulphur bacteria. If the anaerobic sulphur bacteria that promote the sulphur cycle under marine conditions are destroyed by an acid environment, it would appear that syngenetic supergene marcasite is evidence of fresh-water deposition for the containing sediments. Marcasite in marine formations indicates an acidification subsequent to deposition.

Examination of supergene iron sulphides occurring in Tertiary brown coals and certain marine clays in Victoria reveals that the sulphides in the coals are almost invariably marcasite, whereas those in the clays are pyrite. This parallels the observations of Newhouse (1927), but whereas Newhouse considered that the sulphides described by him had been deposited as gels, contemporaneous in origin with the containing rocks, some of the sulphides to be described herein appear to have formed as crystalline substances, after the deposition of the sediments with which they are associated.

IRON SULPHIDES IN BROWN COALS

The sulphur content of Victorian brown coals ranges from about 0.25 per cent up to an exceptional value of 9.2 per cent (Edwards, 1945). Few coals contain more than 5 per cent sulphur, however. The many analyses of bore cores available from the thick seams of the Latrobe Valley area, in southeastern Victoria, show that these coals contain 0.25 to 0.40 per cent sulphur, with a tendency for the sulphur content to increase sharply above this figure near and at the top of the seam. The greater part of the sulphur in the body of the seam is organic sulphur in the coal substance, and sulphate sulphur in the coal water; but the increase in sulphur content at the top of the seam arises from a development of marcasite. This is observed to be so in the Yallourn Open Cut (Yallourn seam: output about 6,000,000 tons a year), and the smaller Yallourn North Open Cut (Latrobe seam: output variable, about 250,000 tons a year). The presence of marcasite nodules in drill cores indicates that it is true also in other parts of this coalfield.

In the open cuts, where there is progressive exposure of the top of the seam, there are numerous nodules of marcasite, up to 3 inches across, replacing fragments of wood. These nodules are concentrated in the surface, and in the uppermost few feet of the coal. Deeper in the seams there is very little marcasite, only a very occasional narrow seamlet in a joint plane.

The marcasite has a steely to grey appearance in the hand specimen, and commonly preserves in detail the texture of the wood (lignite) that it has replaced. In polished sections it shows the characteristic color, anisotropism, and etching behaviour of marcasite. If kept indoors exposed to air, it oxidizes within a few months to a friable mixture of water-

soluble ferrous and ferric sulphates. In the open cuts oxidation is equally rapid, but the soluble sulphates are removed by rain, so that marcasite bodies disappear from any area of coal that has been exposed for a period of years.

Pyrite with Marcasite

In the Maddingley seam, at Bacchus Marsh, which contains about 1.5 per cent sulphur in the lower 50 ft. of the seam, and up to 3.5 per cent sulphur in the upper 30 to 50 ft., there is again a concentration of sulphur as nodules of marcasite near the top of the seam. Some marcasite occurs replacing wood at deeper levels. In addition a little pyrite is present. The pyrite occurs as well developed brassy crystals, up to 2.0 mm across, and is restricted to seams 1 to 3 mm wide, filling transverse fractures in logs of wood. These fractures characteristically are at right angles to the long axes of the logs, across the grain. They are presumably an outcome of shrinkage of the wood, combined with flattening during compaction. Such cross-fractures characterize logs of wood in all brown coal seams, and pyrite, if present, is restricted to such fractures. The development of pyrite in these fractures at Bacchus Marsh may be related to the considerable percolation of water through the coal in that area, the movement of the water being pronounced along such fractures. Logs of wood exposed in the underground workings of the Maddingley mine drip water, whereas the earthy coal enclosing the logs is relatively dry (i.e., its moisture is chiefly inherent moisture). It appears, therefore, that the cellular texture of the logs, their surfaces and the fractures transecting them, provide channelways for surface waters to follow. Where the Maddingley seam was worked in the Bacchus Marsh Colliery (at Parwan), at a depth of about 400 ft., below an impervious cover, the workings were dry, and there was no development of pyrite seams in the cross-fractures of the few logs exposed, though there was a development of marcasite in the dry sands immediately overlying the coal seam, with some nodules of marcasite replacing wood in the upper part of the coal.

At Yallourn the coal as a whole is "dry." Surface waters tend to travel along "mining cracks," and there are no pyrite seams in the cracks in the logs.

It appears that where iron sulphide has formed by replacement of wood, it has the form of marcasite; where it developed in an open space, serving as a water channel, it has the form of pyrite. The pyrite has the characteristic color and isotropism of pyrite in polished sections, shows no relict textures, and responds to a pyrite geo-thermometer (Smith, 1947) in accordance with its low temperature origin.

Composition of the Marcasite

Analyses made by Mr. F. D. Drews (Melbourne Ore Dressing Laboratory, Commonwealth Scientific and Industrial Research Organization) of marcasite samples from the upper surfaces of the Yallourn seam (Yallourn Open Cut), and the Maddingley seam (Lucifer Open Cut), are given in table 1, and show that in both samples the sulphur is present in excess of the amount required for the stoichiometric proportions of FeS_2.

Origin of the Marcasite

Under laboratory conditions marcasite forms from acid solutions, whereas pyrite requires neutral or slightly alkaline solutions (Allen, Crenshaw and Johnston, 1912; Tarr, 1927; Twenhofel, 1932). The water of the brown coal (as distinct from water percolating through it) constitutes from 50 to 70 per cent of the coals under discussion, and is acid, with a pH of about 5.5 to 6.5 (determined by indicator tests), so that during the replacement of wood fragments (lignite) the requisite acid conditions would be present for marcasite formation. The development of marcasite as a replacement of wood, in contrast to seams of pyrite in fractures across the logs, might be explained in terms of relative alkalinity, solutions in-

TABLE 1.—*Composition of supergene marcasite associated with brown coal*

	Yallourn Seam		Maddingley Seam	
	(a)	(b)	(a)	(b)
Fe	40.65	45.83	44.60	46.39
S	48.05	54.17	51.55	53.61
	88.70[1]	100.0	96.15[1]	100.0
Formula	$Fe_{0.972}S_2$		$Fe_{0.996}S_2$	

(a) As analyzed.
(b) Calculated to 100 per cent iron sulphide.
[1] Remainder is carbonaceous matter.

filtrating and reacting with the wood tending to become more acid than those traversing open fractures.

The source of the sulphur is more difficult to explain. The concentration of sulphide sulphur at the tops of the coal seams cannot be attributed to plant processes, such as give rise to sulphur enrichment in the upper layers of peat deposits (Thiessen, 1920), because in almost every occurrence under discussion the coal surface is an erosion surface. The amount of coal eroded varies from a few feet to as much as 100 feet for a particular seam, so that the horizon of the coal surface and of sulphur enrichment is fortuitous. The erosion occurred, moreover, at different periods in different seams. In the Parwan area the Maddingley seam was eroded soon after the deposition of the coal, because the eroded surface of the coal seam is overlain by sands, ligneous clay, and fossiliferous marine clays of middle Tertiary age (Parr, 1942). At Yallourn, by contrast, erosion of the Yallourn and Latrobe seams probably occurred in the late Pliocene or Pleistocene, following the period of block faulting that gave rise to the South Gippsland Highlands, with their still prominent and only slightly dissected scarps. There can be little doubt, therefore, that most of the sulphur in these marcasite nodules was deposited later than, sometimes much later than, the vegetable matter that formed the brown coals.

At Parwan a slightly consolidated sand that immediately overlies the eroded surface of the Maddingley brown coal seam contains a considerable development of marcasite. Numerous marcasite concretions have been encountered in drill cores in the sands immediately above the Morwell and Yallourn seams in the Yallourn-Morwell district. These concretions are from 1 to 3 inches in diameter, and consist of a cement of marcasite around numerous grains of quartz sand. These occurrences suggest that the sulphur in the marcasite nodules, and by inference that in the marcasite in the top of the coal seams, is derived from a source outside the coal seams.

A possible source of sulphur outside the coal seams is ground water. Analyses of river waters and bore water in the Yallourn-Morwell area, kindly supplied by Mr. V. G. Anderson, indicate that present day ground waters in this area do contain sulphur. The surface waters appear richer in sulphur than the bore waters, pointing perhaps to removal of sulphur during the underground passage of the bore waters (table 2).

Of the two, the surface water would be the more likely source of sulphur in the Yallourn-Morwell district, in view of the shallow overburden (generally less than 50 ft.) above the coal, and the fact that the underground waters, though artesian to sub-artesian in character, are sealed below the coal. However, these waters tend to be alkaline, so that while

TABLE 2.—*Composition of surface and underground waters of the Yallourn-Morwell area* (parts per million)

	Surface Waters		Underground Waters				
	1.	2.	3.	4.	5.	6.	7.
Ca	27.8	11.8	14.9	53.9	54.5	20.1	17.5
Mg	11.8	11.8	7.3	13.9	15.3	12.3	12.6
Na	32.8	38.6	95.8	122.4	119.6	102.1	94.2
K	5.9	5.3	15.2	12.2		12.3	12.9
SO_4	10.5	3.3	3.0	1.8	2.1	0.5	0.4
HCO_3	80.5	51.2	170.5	345.0	321.5	222.0	200.1
Cl	80.0	86.6	105.0	132.8	137.0	110.0	104.5
NO_3	0.9	—	—	0.1	tr.	—	—
SiO_2	28.3	20.6	32.8	33.7	27.5	24.4	20.1
Fe_2O_3, Al_2O_3	3.1	4.5	4.3	3.2	3.0	2.0	2.0
pH	7.9	7.4	7.0	7.2	7.2	6.8	6.8

1. Billy's Creek (Morwell Town Supply), (1944)
2. Morwell River (1944)
3. Boola Boola Oil Co. bore, Morwell (1944)
4. Australian Paper Mills Hot Bore, Maryvale (1941)
5. Australian Paper Mills Hot Bore, Maryvale (1945)
6. Hot Spring, South Morwell (1944)
7. Sub-artesian bore, allot, 54, parish of Maryvale (1944)

they are suitable for deposition of pyrite, some modification would be necessary to permit the formation of marcasite.

A second possible source of sulphur would be the sulphates present in water pressed from the coal during compaction. The coal ash commonly carries as much as 20 per cent SO_3, which occurs chiefly as sulphates in the coal water (Edwards 1945). However, at Yallourn and Morwell, such compaction as the coals have undergone occurred before the period of erosion that produced the coal surfaces on which the marcasite concentration developed.

A third source of the sulphur would be from sulphates dissolved in the water that effected the erosion of the coal; a fourth source is the sulphur in the eroded coal.

The process whereby sulphate sulphur is reduced to sulphide is difficult to explain. Reducing conditions exist at the coal surface, and within the coal, while it is under cover. This is established by (1) the continued existence of the marcasite nodules as long as the coal is covered, and (2) the avidity with which freshly exposed brown coal absorbs oxygen. Some writers consider that coal substance itself is capable of bringing about a reduction from sulphate to sulphide. However, in the laboratory and in industrial plants, reduction of sulphate to sulphide, in the absence of hydrogen sulphide, is only achieved by heating the sulphate with carbon to temperatures in the region of 700° C. (Ginter, 1938; Bastin, 1926). Allen and his associates (1912) failed to reduce ferrous sulphate in the presence of starch at 300° C., provided hydrogen sulphide was absent. A little sulphide sulphur is present throughout the brown coal, presumably as minute particles of marcasite, similar to those described by Thiessen (1920) in black coals, but it is doubtful if this sulphide sulphur was sufficient, or in the right state, to reduce sulphates in waters penetrating to the coal surface, in view of the abundance of sulphate sulphur relative to sulphide sulphur

throughout the brown coal seams as a whole. It appears, therefore, that the formation of marcasite must be attributed to some factor other than simple reduction of sulphates on coming into contact with coal substance.

Various anaerobic bacteria that inhabit peat swamps can reduce sulphates to hydrogen sulphide which will react with any ferrous salts available to form ferrous sulphide (Thiessen, 1920), and it is possible that such bacteria were active at the surface of the coal when it was undergoing erosion, or just subsequently. The surface layer of the coal seam, as now preserved, consists of a foot or more or black ligneous muck or ooze. The conversion of the brown coal, with its abundant residual plant structures to this black muck, during erosion or subsequently, involved a considerable oxidation and decomposition of the coal constituents; but it is clear that reducing conditions existed in the black muck or were subsequently re-established there. This suggests the possibility that anaerobic bacteria were present, and if these included some variety of sulphur bacteria, the concentration of sulphides at the surface of the coal—the base of the ooze—can be explained. The sulphur could be derived from the degraded coal, by a process of repeated concentration during erosion, or from the waters covering the black muck, or both.

The delicate replacement of the wood structures by the marcasite calls for colonies of bacteria living in the cells of the wood fragments, developing hydrogen sulphide within the cells, and for the fixation of the sulphide sulphur as marcasite by reaction with iron salts, probably iron humates, impregnating, the wood, or diffusing into the water filling the cells. It is difficult to imagine how otherwise the replacement, including the solution of the humified wood substance, could proceed. The filling of cells in wood by iron sulphide has been achieved by Stocks (1902) in much this same way, by exposing wood to the action of reducing anaerobic bacteria in the presence of iron hydrate and calcium sulphate solutions.

The marcasite concretions in the sands overlying the coal seams would, on this view, be younger than the marcasite nodules in the coal which the sands have re-buried. In the Morwell area, the marcasite concretions occur chiefly within a few feet of the top of the coal seams in sands saturated with water. In the Bacchus Marsh Colliery, at Parwan, the marcasite occurs in dry sands immediately overlying the seam. These sands are sealed from present day surface waters by marine clays, a thin impure coal seam (really a ligneous clay), thin limestone beds, and a cover of basalt. The impression is that such marcasite is derived from percolating waters, but the conditions of deposition cannot be established.

Association of Marcasite with Brown Coals

The association of marcasite, whether syngenetic or epigenetic, with brown coal is not peculiar to Victorian brown coals, but appears to be characteristic of brown coals in general. So also is their acid nature, which is presumably inherent from the acid conditions prevailing in the original peat swamps. Marcasite bodies similar to those at Yallourn characterize the relatively sulphur-rich brown coal at Moorlands, South Australia, the only South Australian brown coal deposit sufficiently developed to establish this point. They are also a feature of the brown coal at Bovey Tracey, in Devon, England. Both these coals give acid reactions. Marcasite replacements of wood are a feature of the Lower Rhine brown coals, which have a pH of 6.0 to 6.5; and there is an abundant development of marcasite, partly as nodules replacing wood, and partly as spherical concretions of radiating crystals, up to 3 inches in diameter, in the upper few feet of the Treue seam at Helmstedt, Brunswick, Germany, where the pH of the coal is about 6. Published data on the occurrence of iron sulphides

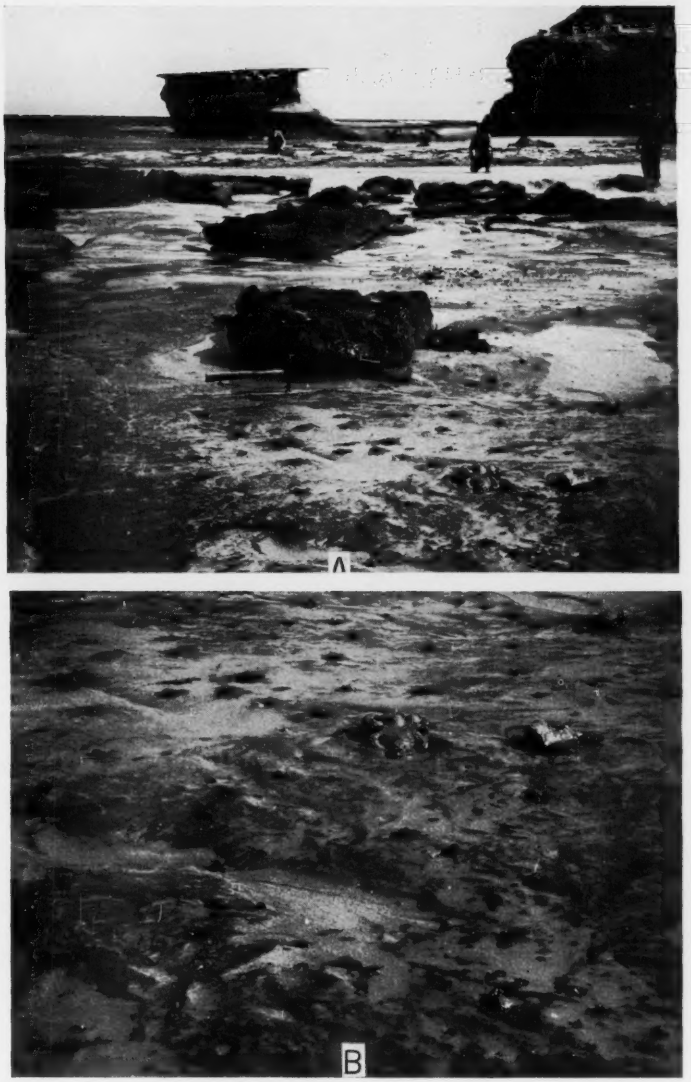

FIG. 1.—

A. Wave-cut platform on the northeast side of Bird Rock, 1½ miles southwest of Torquay, Victoria, showing the limestone reef, and the pyritic Ancilla clay.
B. Pyrite nodules and crystals protruding from the Ancilla clay.

(F. A. Singleton; photographs)

in German brown coals does not often distinguish between pyrite and marcasite, but Erdmann (1927) indicates that marcasite is the prevailing form. There is a similar lack of precision about the form of the iron sulphides in the brown coals of North Dakota. Hares (1928) describes marcasite replacements of wood, with the grain structure preserved, and ball-shaped masses, rosettes and feather-like growths of marcasite in brown coals of both the Fort Union and the Lance formations of the Mammoth coalfield in North Dakota. One of the present authors has observed marcasite replacements of wood in brown coals near Velva and Hazen, North Dakota.

IRON SULPHIDES IN MARINE CLAYS

In contrast to the predominance of marcasite in the brown coals, is the occurrence of pyritic nodules and crystals, and pyritic casts of marine organisms in certain Tertiary marine clays exposed along the coast of south central Victoria at Torquay, Point Addis, and Curlewis.

Torquay

Torquay is about 50 miles southwest of Melbourne, and the occurrence under discussion is in Half Moon Bay, on the northeast side of Bird Rock. The pyritic nodules and crystals occur on the wave-cut platform, in the upper part of a flat-dipping (to the north) bed of dark blue-grey clay (the lower Ancilla clay) of Janjukian (upper Oligocene to lower Miocene) age, between two curving reefs of hard limestone (fig. 1, A). The outcrop is on the northeast side of a denuded dome, and is stratigraphically above the Bird Rock Cap horizon, and below a hard band at the base of a Septarian limestone horizon. The nodules are widely spaced (fig. 1, B) and are readily detached from the clay, which contains no disseminated pyrite in their vicinity. They are uniformly distributed through a thickness of about 13 ft. of clay (Point Addis No. 3 bore), and their distribution shows no obvious relation to bedding or jointing. The clay is sparsely fossiliferous, and a number of the organisms have been infilled or replaced by pyrite, when the shape of the organism has determined the shape of the nodule. The calcareous shells or structures of the organisms are frequently preserved (fig. 2, I) indicating that the pyrite was deposited under neutral or alkaline conditions. The nodules (fig. 2, A–H) are up to 10 cm across. They commonly consist of aggregates of small crystals, which rarely exceed 1.0 mm in size. Some show radiating and stalactitic growths; some consist of attached and incompletely developed crystals. Others are compound crystals in parallel growth. Isolated individual crystals are uncommon, but isolated compound crystals are relatively abundant (fig. 3). The individual crystals may be up to 15 mm across.

EXPLANATION OF FIGURE 2

A and H. Organic structures (indeterminate) replaced and encrusted with minute pyrite crystals (both natural size). Torquay.
B and D. Elongated nodules consisting of coarser crystals (both natural size). Torquay.
C. Irregular nodule of coarse and fine crystals of pyrite (natural size). Torquay.
E, F and G. Rounded nodules with small and medium size crystals. (all ×2). Torquay.
I. Zoarium of a bryozoan infilled with and partly replaced by pyrite (×1½). Torquay.
J. Internal cast of pyrite, with a remnant of the calcareous shell preserved, in a pyritic matrix (natural size). Curlewis.
K. Original calcareous shell in pyritic matrix. The shell is partly disrupted by the growth of the pyritic filling (natural size). Curlewis.
L. Internal cast in pyrite of the pteropod *Hyalocylix* (?) *annulata* (Tate) (×8). Point Addis.
M. Pyritic internal cast of a spatangoid echinoid (natural size). Point Addis.
N. Side and apical view of a pyritic internal cast of a cowrie, in a matrix of pyrite. (natural size). Curlewis.

(M. Johnston; photographs)

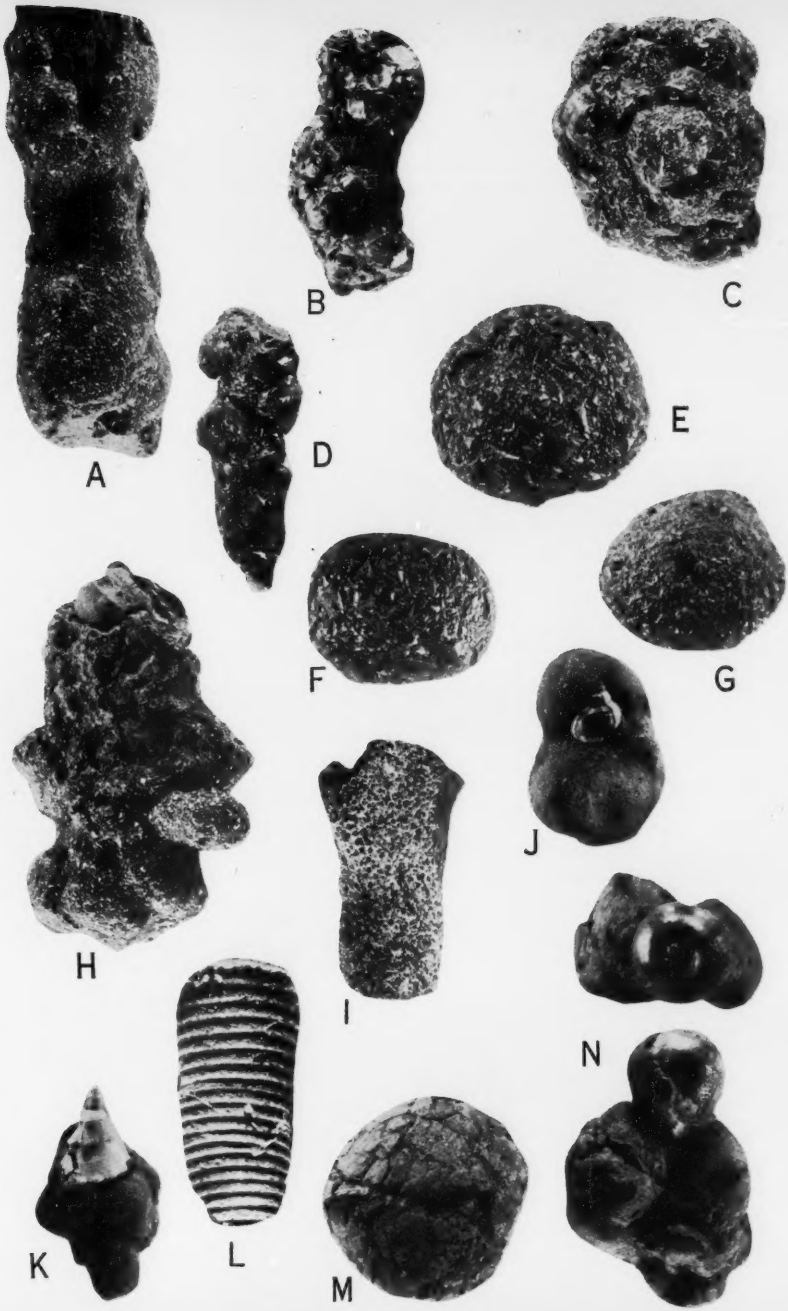

FIG. 2.—Pyritic replacements and nodules from Torquay, Point Addis, and Curlewis. A–I from Torquay.

The majority of the crystals consist of cubes modified by octahedra, and in some the two forms are equally developed (fig. 3, A–F). Pyritohedral faces are practically absent. In some crystals the central areas of the cube and octahedron faces are depressed, and consist of parallel slightly stepped faces (fig. 3 A, B). In others, the central areas of these faces are occupied by groups of small outgrowing crystals (fig. 3, A, B, C, E, F). Commonly the octahedral faces possess "scale-like" surfaces resulting from growths of small sub-parallel piles of minute octahedra upon them. Less commonly similar growths occur on the cube faces (fig. 3, D). Occasional crystals are interpentration twins (fig. 3, B, F). Striation of faces is imperfectly shown on some of the concave surfaces of the crystal growths (fig. 3, K). The larger crystals commonly have concave cube faces (fig. 3, G–K) whereas concave octahedral faces are unusual. In many the concavities are partly infilled with growths of small irregular crystals (fig. 3, J, K). In the ultimate stage, such growths form nodules.

Pyrite is not confined to the Ancilla clay. Thin veins of fine-grained pyrite infill concentric and radial cracks in some of the septaria of the immediately overlying Septarian limestone, and small clusters of pyrite crystals and narrow veins of pyrite occur in other calcareous beds at this locality.

The pyrite is pale brassy-yellow. occasionally with an iridescent tarnish. The specific gravity of the crystals averages 4.95, that of the nodules 4.80. The lower value for the nodules is due to cracks, cavities, and inclusions of carbonate. Polished surfaces reveal that pyrite is the only sulphide mineral present. No marcasite could be detected, though occasional inclusions of carbonate were noted. A number of the crystals show lineage textures, and tests with a pyrite geo-thermometer (Smith, 1947) are in accord with its low temperature origin.

Composition of the Pyrite

Chemical analyses made by Mr. F. D. Drews (Melbourne Ore Dressing Laboratory, Commonwealth Scientific and Industrial Research Organization) of a sample of pyrite cubes, and a sample of nodules are shown in table 3. Recalculation to 100 per cent on a pure pyrite basis reveals that in each sample the iron content is in excess of that required for the ideal composition FeS_2, the reverse of the situation found with the marcasite.

This pyrite does not generally oxidize in storage, but a few crystals have decomposed to a light grey powder after 10 years in storage, and occasional nodules show slight decomposition to ferrous and ferric sulphates along cracks. The pyrite in the Ancilla clay is protected from weathering by periodic wetting by sea water, but pyritic nodules in other beds in this locality, exposed to normal atmospheric weathering at Bird Rock, are converted to limonite nodules, or to jarosite, and less frequently, to copiapite. About 3 miles to the southwest, in the upper part of the cliff section, a former deposit of pyritic (or marcasitic?) nodules in Anglesean carbonaceous sandy clays is completely converted to jarosite and copiapite, and has been worked as a source of ochre.

EXPLANATION OF FIGURE 3

A–F. Combination and compound crystals (all ×3).
G–L. Skeletal crystals with concave faces, striated faces, and groups of smaller crystals encrusting the concave faces.
G. Cube and octahedral faces concave (×2½).
H and I. Concave cube faces (×3).
J and K. Concave, striated and encrusted cube faces (×2).
L. Concave octahedral face (in center) with three concave cube faces (×2).

(M. Johnston; photographs)

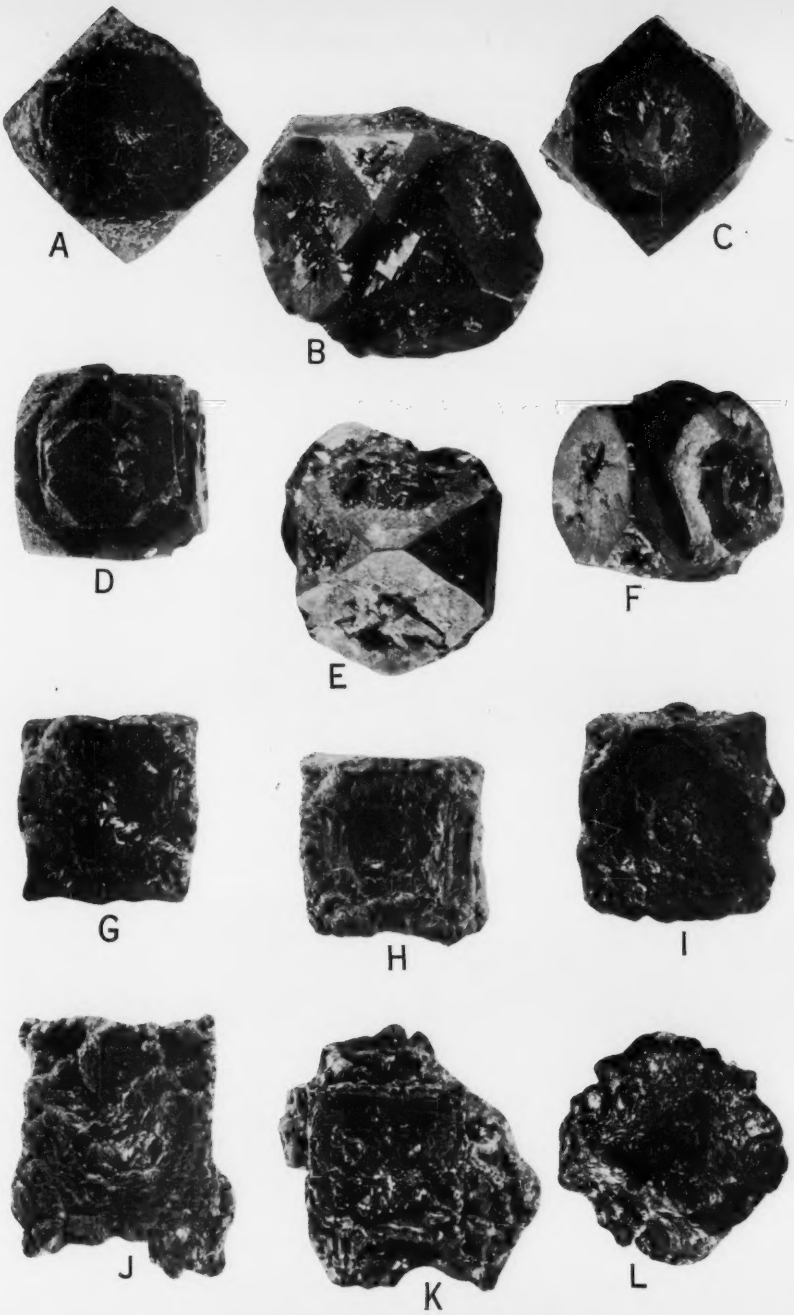

FIG 3 —Pyrite crystals from the Ancilla clay, Torquay.

TABLE 3.—*Composition of supergene pyrite in marine clays at Torquay, Victoria*

	Cubes		Nodules	
	(a)	(b)	(a)	(b)
Fe	46.45	47.35	45.80	47.26
S	51.65	52.65	51.10	52.74
Insol.	0.76	—	1.80	—
	98.85[1]	100.0	98.70[1]	100.0
Formula	$Fe_{1.013}S_2$		$Fe_{1.03}S_2$	

(a) As analysed.
(b) Recalculated to 100 percent pyrite.
[1] Deficiency is carbonate.

Point Addis

Sparsely occurring pyritic nodules and pyritic replacements of fossils occur in Anglesean (upper Oligocene) carbonaceous sandy clays north of Point Addis, which is midway between Torquay and Anglesea, about 55 miles southwest of Melbourne. The pyritic replacements occur about 15 to 20 ft. above sea level, at the base of cliffs up to 200 ft. high, cut in the Anglesean sediments. The replacements comprise internal casts of the pteropod *Hyalocylix (?) annulata* (Tate) (fig. 2, L), and spatangoid echinoids (fig. 2, M). The calcareous shells of these fossils have not been observed although tests of the lituolid foraminifera *Cyclammina*, and sharks' teeth (*Odontaspis*) are preserved in the clay. Pyritic casts of *Cyclammina* are also present in the bed. Polished sections prove that the material is pyrite, without any admixed marcasite.

The beds at this locality have been broadly warped by post-Miocene (probably late Pliocene) earth-movements, but are not otherwise altered.

Curlewis

At Curlewis, on the northern shore of the Bellarine Peninsula (Port Phillip Bay), about 10 miles east of Geelong, occasional pyritic nodules occur in a gently warped, bluish-black marine clay of Balcombian (Miocene) age, exposed between high and low tide level at locality Ad. 14 of Hall and Pritchard (1894). The nodules commonly enclose casts or moulds of fossil shells (fig. 2, J, K, N). Some of the fossils have been infilled by pyrite with preservation or partial preservation of the calcareous shell (fig. 2, K, J), providing evidence that the pyrite was deposited under neutral or alkaline conditions, either contemporaneously with the clay, or subsequently. Hall and Pritchard (1894) record that some shells have been crushed in contact with the pyrite nodules. Others observed by the present authors have been fractured by the growth of the pyritic material within the shells. One univalve shell unfilled with pyrite has been disrupted along its sutures, and along cracks developed obliquely to the sutures.

Polished sections reveal that the sulphide is pyrite, without any trace of marcasite, and the nodules show no tendency to oxidize, even after prolonged storage. Hall and Pritchard (1894) noted that some nodules washed out of the clay had developed a crust of limonite, and that oxidation of the pyrite in this way in situ had stained the clay yellow in the immediate vicinity of the weathered nodule.

Origin of the Pyrite

The blue-black to blue-grey color of the clays indicates that these pyritic

nodules developed under reducing conditions, and their association with fossils suggests that they were localized by dead organisms, possibly in a marine environment. The preservation of the calcareous shells of the organisms suggests alkaline or neutral conditions; and the fact that weathering of the nodules in exposed clays leads to the precipitation of nodules of limonite, indicates that the clays themselves exert a neutralizing effect on free acids. The volume of sulphur in a nodule would far exceed the amount of sulphur present in the body of a dead organism, and if Bastin (1926) is correct in stating that "there is no evidence for considering dead organic matter capable of reducing sulphates at ordinary temperatures," it seems necessary to invoke the aid of anaerobic sulphur bacteria to explain the concentration of sulphur in the nodules, and in the internal fillings to the shells. The reducing atmosphere, and the apparently neutral or alkaline conditions, are in accord with Galliher's (1933) account of the conditions required to permit the activities of anaerobic sulphur bacteria under marine conditions. Acid conditions destroy the anaerobic bacteria that promote the sulphur cycle under marine conditions.

If the pyritic nodules developed concomitantly with the deposition of the clays, there is thus no difficulty in accounting for their formation. If they formed subsequently the growth of such large crystals would have required a considerable diffusion of sulphates through the clays, implying that the sulphides crystallized before compaction had greatly reduced the water content of the clays. The infilling of the shells is also more difficult to explain.

It is not possible to determine whether the iron sulphide was precipitated as a gel that crystallized subsequently, or whether it crystallized directly. The complex forms of the crystals suggest growth in stages, with limited supplies of sulphur at any state, pointing to direct crystallization. The casts of the shells give no indication of shrinkage from the shell, such as might be expected from a crystallized gel. The significant fact, however, is that the neutral or alkaline character of the environment remained unchanged during crystallization.

CONCLUSION

The widespread development of marcasite in association with the acid environment of brown coals is in striking contrast to the development of pyrite in marine clays that provided a neutral or alkaline environment, and parallels the precipitation of the two forms of ferrous sulphide under laboratory conditions. It is concluded, therefore, that the form of ferrous sulphide found in supergene deposits may be used as an indicator of the acidity or otherwise of the conditions attending its deposition, even though the factors responsible for the deposition of the iron sulphide may not be clear.

The replacement of humified wood (lignite) by marcasite with perfect preservation of the wood structures seems best accounted for by the action of colonies of anaerobic bacteria living in the wood cells, a process that can be demonstrated experimentally. Similarly, the deposition of pyrite as the filling of marine shells, of complicated form, seems best accounted for by the action of colonies of anaerobic bacteria living on the body of the dead organism.

Anaerobic sulphur bacteria can flourish in the acid environment of peat swamps, but the anaerobic bacteria that promote the sulphur cycle under marine conditions require an alkaline or neutral environment, and are destroyed by acid conditions (Galliher, 1933). This suggests that supergene marcasite does not form in a marine environment, so that syngenetic marcasite is evidence of fresh-water deposition. The presence of marcasite in marine formations would, therefore, indicate an acidification subsequent to deposition or during lithification.

ACKNOWLEDGMENT

We are indebted to Mr. F. S. Colliver for the loan of his collection of crystals and nodules of pyrite from the clay beds at Torquay, and of replaced shells from Curlewis.

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FIGS. 1–5, MARCH, 1951

LONGITUDINAL RIPPLE MARKS IN MUD AND SAND

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ABSTRACT

In the shallow Waddensea, off the coasts of Holland, Germany, and Denmark, the bottom is uncovered during ordinary low tide. Observations and measurements of ripple marks in both mud and sand, correlated with observations of environmental factors such as wind, waves, and currents, indicate that ripple marks in mud generally are symmetrical and are formed with their crests parallel to the direction of movement of currents. Normal, asymmetrical current ripples in sand patches are formed with their crests perpendicular to the crests of the symmetrical ripple marks in the surrounding mud. In addition to these transverse ripple marks, longitudinal ripple marks also are formed in sand, in some places in combination with the transverse marks.

It is suggested that fossilized longitudinal ripple marks may be used as an indication of extremely shallow water in which currents of long duration acted in a constant direction. Moreover, they may provide indications of the orientation of nearby banks or shores.

INTRODUCTION

A shallow sea, known in Dutch as the "Waddenzee," exists off the coasts of Holland, Germany, and Denmark. The sea is a tidal flat area, separated from the North Sea by a series of small islands, and its bottom is uncovered for the greatest part at ordinary low tide. The writer investigated the marine geology in the Dutch part of this area, and found that some kinds of ripple marks, hitherto little known, had a common occurrence. The following types are discussed in this article; longitudinal ripple marks in mud, wave-formed ripple marks in mud, and longitudinal ripple marks in sand.

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THE WADDENSEA MUD

The mud of the tidal flats of the Waddensea is often deposited in the form of small pellets which are extruded by invertebrate organisms. Foremost among the producers of these fecal pellets are *Cardium edule* and *Mytilus edulis*. It might be thought that the formation of ripple marks in mud is the consequence of the granular state of the sediment, which is deposited as a mixture of sand grains and fecal pellets; however, the mud is distinctly granular only in a few places where rippled areas are found. Most often the pelitic fraction of the mud is deposited as much finer elements than these fecal pellets.

Figure 1, A, shows a grain size analysis of a rippled mud; the sand content is relatively large, but no larger than in other Waddensea muds. Despite the sand, in their natural wet state these muds present all the properties typical of mud, such as plasticity, cohesion, and thixotropy.

RIPPLE MARKS IN MUD

Previous Investigations

As far as the writer is aware, the existence of ripple marks in mud has been given very little attention. Richter (1926) mentioned their common occurrence in the German Waddensea, but gave no single detail. Trusheim (1929) reported ripples which he interpreted as the result

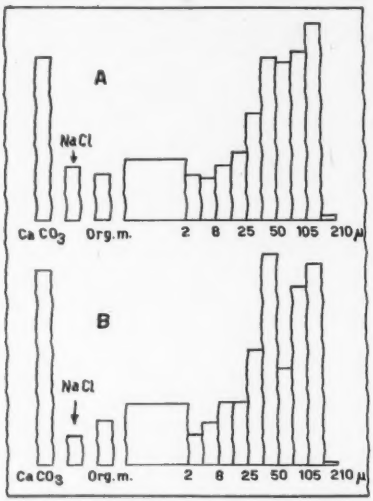

FIG. 1A.—Grain size analysis of mud from Waddensea bottom, SW. of Nes (Ameland), in which longitudinal mud ripples were present November 17, 1949.

FIG. 1B.—Grain size analysis of mud from bank of Wester Scheldt near Bath (Zealand), in which wave ripple marks were present October 26, 1949.

Table accompanying fig. 1A

	Percentage		Percentage
$CaCO_3$	10.7	16– 25μ	4.5
NaCl	3.5	25– 35μ	7.0
Organic matter	3.0	35– 50μ	10.7
<2μ	16.6	50– 75μ	10.4
2– 4μ	2.9	75–105μ	11.1
4– 8μ	2.7	105–150μ	12.9
8–16μ	3.6	150–210μ	0.4

Table accompanying fig. 1B

	Percentage		Percentage
$CaCO_3$	12.7	16– 25μ	4.2
NaCl	1.8	25– 35μ	7.4
Organic matter	2.9	35– 50μ	13.9
<2μ	17.0	50– 75μ	6.3
2– 4μ	1.9	75–105μ	11.6
4– 8μ	2.7	105–150μ	13.2
8–16μ	4.2	150–210μ	0.2

of wave action. From his description and photographs, however, the conclusion appears to be warranted that these ripple marks are of the type called "longitudinal ripple marks" in the present paper. Frebold (1928) found symmetrical ripple marks with rounded crests and troughs in an erratic boulder of Gothlandian Graptolite shale from Rügen (Germany). Other writers have expressed the view that mud particles are too small and too cohesive to roll over the bottom and thus to form ripple marks in the manner of sand grains. The validity of this view is not known to the present author, because the turbidity of the water has prevented direct observation of the actual process by which the mud ripples are formed. It might differ essentially from the process which gives rise to ripple marks in sand.

Longitudinal Ripple Marks in Mud

The most common kind of ripple marks in the Waddensea mud is symmetrical in transverse profile (fig. 2), and the troughs and crests are about equally rounded. The crests are remarkably straight and commonly are parallel for more than ten meters (fig. 3). The observed wave lengths (distances between crests) range from 1.5 to 6 cm., with an average value of about 5 cm. The average ripple index (ratio of the wavelength to the amplitude) ranges from 10 to 12.

For the following reasons the ripple marks are thought to be caused by the combined effects of tidal currents and waves:

1.—Observations of the wind and waves moving over these ripple marks, show that they have a random orientation with respect to the orientation of surface wave crests, and that the two do not become aligned even after long periods of strong and constant winds.

2.—The ripple marks are exactly parallel to the dominant directions of movement of the tidal currents.

3.—Normal asymmetrical transverse current ripple marks are produced on patches of sand which occur in the mud flats. The crests of these ripple marks are

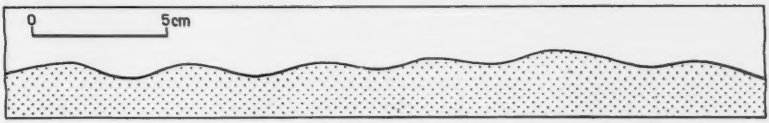

FIG. 2.—Transverse section of longitudinal mud ripple marks, mud flat SW. of Nes (Ameland), July 20, 1948.

always perpendicular to the crests of the symmetrical ripple marks in the surrounding mud (fig. 4).

4.—The ripple marks in mud curve around obstructions such as clumps of mussels on the mud flats. The curves resemble flow lines (stream lines), which suggests that the ripple marks are longitudinal.

5.—Experiments carried out in the laboratory of experimental geology at Leyden University, show that in normal circumstances flow of water in itself does not produce regular systems of longitudinal ripple marks in mud. Sediment was collected from ripple marked mud surfaces in the Waddensea and sent directly to Leyden. There it was allowed to settle in water so as to obtain about the same state of surface cohesion as in the original environment. (Fresh) water was made to flow over this mud with various velocities and depths and for various periods of time. But, whatever the circumstances, only irregularly spaced, very indistinct flat ridges and shallow troughs were obtained, extending only over short distances.

6.—The ripple marks disappear in very calm weather and reappear in the course of a few days when strong winds set in.

7.—As in the case of ordinary wave-formed ripple marks, the distances between the crests of the longitudinal ripples are dependent on the dimensions of the wind waves and on the water depth, as will be illustrated in more detail by the following example.

On most mud flats only one system of ripple marks is present. Yet the tidal cur-

FIG. 3.—Longitudinal ripples in mud, SW. of Nes (Ameland), July 20, 1948.

FIG. 4.—Longitudinal ripples in mud (front) with crests standing perpendicular to those of transverse ripples in sand (background), SW. of Nes (Ameland), July 9, 1947.

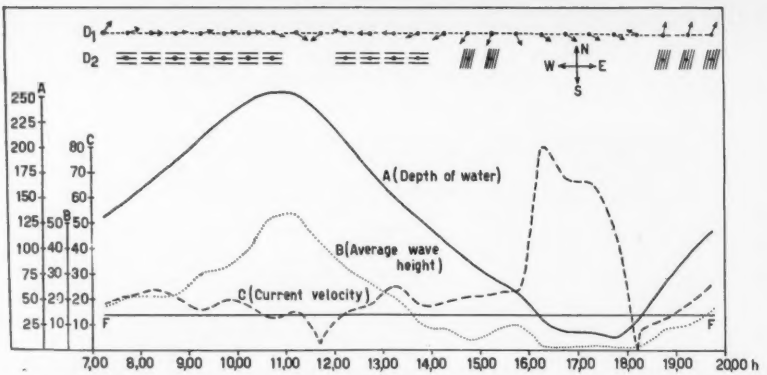

FIG. 5A.—Data on movement of water at locality A (see map, fig. 5B), above bottom of ebb gully on August 20, 1948.
Horizontal: Time
Vertical: A) Depth of water (cm.)
B) Average wave height (cm.)
C) Current velocity at surface of water (cm./sec.)
FF) (Horizontal line) Height of bank above bottom of gully
Directions: D1) Directions of current at 40 cm. below surface of water
D2) Directions of crests of longitudinal mud ripples present on August 8, 1949 on bank of gully near A.

N. B. The wind velocities on August 20, 1948, at 1 m. above the water surface, measured every 30 minutes, were from 07 15 until 19 45 successively: 5.0, 5.0, 4.7, 4.5, 4.7, 5.9, 8.1, 6.1, 7.0, 5.6, 6.6, 4.4, 4.3, 4.4, 3.3, 5.0, 6.1, 6.1, 3.9, 6.6, 5.6, 4.8, 5.1, 4.5, 5.0. In August 1949 the average velocity was ca. 7.0 m./sec.

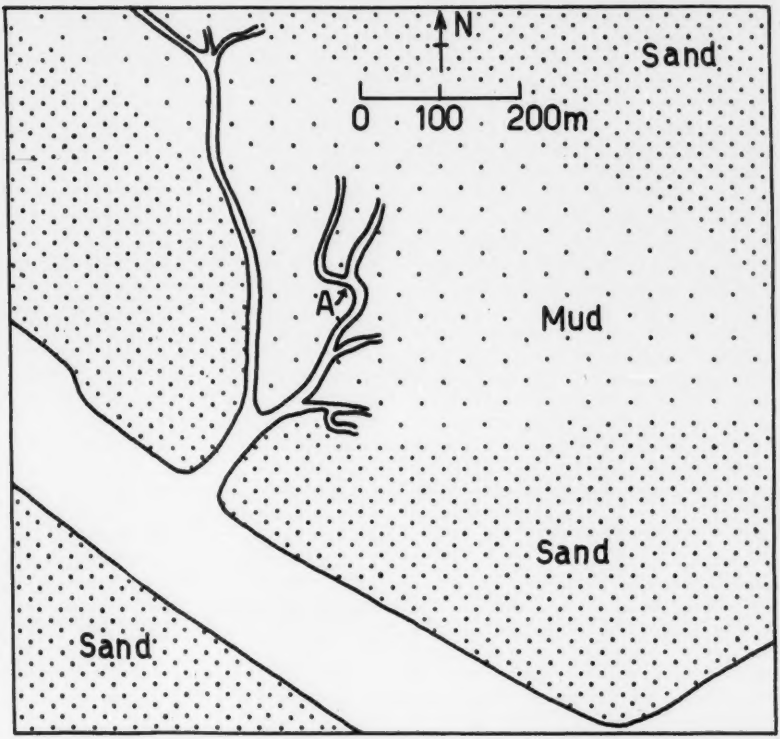

FIG. 5B.—Observations on movement of water and on longitudinal mud ripple marks were made at locality A. The nearest coast is 800 m. north of locality A, and trends approximately east-west.

rents do not move all the time in one direction or its opposite only. The periods during which the different directions of movement are followed, however, vary considerably and at most places movement in only one direction persists long enough to produce lasting effects on the bottom relief. Nevertheless, in some places these time conditions are more favorable and two or even three intersecting systems of ripple marks, each system with a different wave length and orientation, may be found. There is such a place near the center of a gully system which debouches in a main channel SW. of Nes (Ameland). The following systems were measured on the bank of one of these gullies (located at "A" on fig. 5B), on August 8, 1949:

1.—Distinct ripple marks, oriented east-west, average wave length 4.85 cm.
2.—Distinct ripple marks, oriented N. 10° E., average wave length 2.63 cm.
3.—Indistinct ripple marks of smaller wave length, oriented parallel to the gully.

The first system generally is present in a large area (the whole of the map, fig.

5B). The second system has only local importance, in the area of the gullies, and the last system is limited to the slopes of the bank of the gully.

Currents had been observed at locality A approximately a year earlier on August 20, 1948. In the course of this year the gully system became modified—meanders were displaced, some minor gullies were silted up and new ones cut at other places. However, no changes occurred at A, nor were the modifications elsewhere sufficient to cause any important difference between the currents on August 20, 1948, and those on August 8, 1949. The tides on both dates were about equally strong; the average wind velocities were a little lower on the earlier day, but the wind directions were the same, SW. to W.S.W. Observations of the water movements on August 20, 1948, therefore, can be used to infer the conditions of origin of the ripple marks measured on August 8, 1949 (fig. 5, A and B).

From the graph, fig. 5A, the following may be gathered:

When the water level sinks below the surface of the neighboring flats (ca. 16 00) the current velocity jumps to high values. Subsequently the remaining water, which is drained by the gully from the uncovered flats towards the main channel, keeps the level in the gully comparatively high, as is apparent from the change in slope of the curve showing the water level (16 15 to 17 45). The level in the main channel sinks much deeper and the difference between high tide and low tide, which is 240 cm. in the gully is correspondingly greater in the channel. The steep slope of the surface of the water between upstream parts of the gully and the channel causes the high current velocities in the gully.

Subsequently the level in the channel rises again, causing a rapid decrease of this slope in the gully. The level at the locality of the observations starts to rise somewhat later (17 45), and the downstream current velocity drops quickly. At 18 10 the slope at this spot is reduced to zero and so is the current velocity. Then the flood current sets in.

With the flood the water soon overflows the banks of the gully and the current loses its dependence on the course of the gully, and changes to the general direction of the (relatively low lying) area in which the whole gully system is incised. This direction is maintained sufficiently long to cause the formation of a system of mud ripples, the N. 10° E. system. After a certain lapse of time the more elevated sand flats (in which no important gullies are present) are also well under water. The whole area is inundated and the current changes its direction once more and becomes more or less parallel to the nearest coast. This is the north coast, 800 meters distant and running approximately E-W. This direction is the dominating direction N. 90° E., corresponding to the bearing of the large majority of the mud ripples and is maintained until high tide. It should be borne in mind that the direction of the surface currents is always more or less influenced by the wind, and that the bottom currents correspond still more closely with the directions of the ripple marks.

From high tide to low tide the currents follow the same directions in the opposite sense and in the reversed order of succession.

As to the waves, it is seen from the graph that there is a close correlation between the depth and the average wave height. The wave height was practically constant, due to the small length of fetch, only when the water level was below the banks of the gully.

By comparing the depths with the current directions, which coincide with the directions of the ripple marks, the following correlation can be made:

Depth of water	Height of waves	Wave length of mud ripple marks
30– 90 cm.	5–17 cm.	2.63 cm.
113–213 cm.	21–53 cm.	4.85 cm.

This relation between increase of the wave length of the ripple marks and increase of the depth of the water was found to be the same in the Waddensea, wherever observations were made. Probably it is valid only to a certain depth, below which the wave length of the ripple marks may decrease with increase in depth.

The relation of ripple mark formation to sedimentation and erosion was investigated by observing the same ripple marks during several tidal cycles. The original position of the rippled surface was marked by placing brass rods, 1 mm. thick, vertically in the mud, so that the top ends just touched the surface. Some rods were placed in the crests, others in troughs or on the slopes of the ripple marks. Limited observations obtained by this procedure suggest the following conclusions:

1.—Ripples may preserve their exact position during several successive tidal oscillations.

2.—Deposition of mud often effaces ripple marked surfaces.

3.—Erosion of mud frequently is accompanied by the formation of new ripple marks.

Therefore, the ripple marks seem to be the products of erosion rather than of sedimentation. This is not surprising, because the ripple marks are formed only when wave action is sufficiently strong, and wave action always stirs up much of the superficial mud. However, it is possible that longitudinal ripple marks in mud may be formed also in those rare places where wave action does not prevent the deposition of mud.

In this connection it is noteworthy that longitudinal ripple marks are not found where *Zostera* forms vegetation covers, even though there are strong tidal currents and vigorous winds. *Zostera* tends to suppress stirring of the bottom, by fixing the mud with its roots and by covering the surface with its leaves, and the water remains relatively clear.

Wave-formed Ripple Marks in Mud

Purely wave-formed ripple marks in mud are much less common than the ripples described above. Their crests are not as straight and parallel, nor as long, as the crests of longitudinal ripple marks. Moreover, the crests usually are less rounded. A grain size analysis of mud taken from wave-formed ripple marks in the Wester Scheldt near Bath (Zealand) is given in fig. 1B. The ripple marks, which had an average wave length of 5.5 cm, were found in a gap in a *Spartina* vegetation cover. The place had been inundated by a high tide during a strong gale. Current action probably had no influence because the spot was situated in a dead corner between three dikes, and because the *Spartina* stood high and dense. Moreover, the direction of the crests was normal to that of the wind, and normal to the direction of the only possible (and very weak) currents.

LONGITUDINAL RIPPLE MARKS IN SAND

The crests of the longitudinal ripple marks in sand usually are much shorter than those in mud, at most a few meters. In other respects the ripple marks in mud and sand have much the same appearance. Longitudinal ripple marks in sand usually are found in combination with the normal, asymmetrical, current-formed ripple marks, to which they are always perpendicular. At times almost all sand flats in the Waddensea are covered by a mosaic pattern of patches, in which either longitudinal ripple marks or transverse ones dominate. If the development of both kinds of ripple marks is about equal at a given spot, one not infrequently observes a thickening of the crests of the longitudinal ripple marks on the crests of the transverse ripple marks.

No observations have been made relating the distance between the crests of these ripple marks to the depth of the water, or the amplitude of the water surface waves. At present it is not known if

the ripple marks are related to the lengthwise current ridges described by Evans (1949).

Use of Longitudinal Ripple Marks in Studies of Paleogeography

"Fossilized" longitudinal ripple marks were observed locally in the Waddensea where erosion of the uppermost sediments to a depth of a few centimeters, had laid bare old bedding planes. The ripples in Graptolite shale mentioned by Frebold, possibly are an instance of their occurrence in old, lithified sediments.

Their presence may be used as an indication of shallow water in which currents of long duration acted in a constant direction. Moreover, they may provide indications of the orientation of near-by banks or shores.

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REVIEWS

Provenance, Transport and Deposition of Rhine Sediments, A Heavy Mineral Study on River Sands from the Drainage Area of the Rhine by Tj. H. van Andel. Pp. X+129, 45 figs. incl. 3 maps, and 28 tables. H. Veenman & Zonen, Wageningen (Netherlands) 1950. For sale by Geological Laboratory of the Agricultural University, Wageningen, Netherlands for D.fl. 4.50 (about $1.40).

The very existence of a large part of Holland is a result of sedimentation by the Rhine. It is little wonder then that the Dutch geologists have a particular interest in that river system, as Edelman has made clear in numerous publications. The present work does not treat of the geology of Holland, nor of the past importance of the Rhine, but it goes far beyond any previous work in the study of its present sediments and their heavy minerals. Beyond its obvious importance for Dutch geology, the report should be of interest for anyone who wants to know how to investigate river sediments with a view to using the heavy mineral data in whatever manner, and likewise for one who is dealing in problems of river transport—sorting, size variations, abrasion, influence of tributaries, and similar aspects.

Apart from a summary and references at the end, the report consists of six chapters, entitled respectively: Introduction; History, Principles and Methods of Heavy Mineral Research in the Netherlands; Causes and Effects of Granular Variation; Description of Minerals; Mineral Associations of the Supply Areas; and, Mineral Associations of the Rhine.

Chapter I (4 pp.) states the chief purposes, and the methods of work (mostly methods used previously).

Chapter II (13 pp.) gives first a concise summary of past heavy mineral work in the Netherlands, and elsewhere by the "Dutch school." Second, there is a brief discussion of the various terms used. For example, Edelman's 1938 definition of his term "sedimentary petrological province" is quoted: "A sedimentary petrological province is made up by a group of sediments, which constitute a natural unity by age, origin, and distribution." A "mineral association" has lower rank; it is "The combination of minerals by which a sediment is characterized." A term widely used by Andel is "granular variations," which are variations in heavy mineral content caused by differences in grain size and by other factors spoken of below. Finally, the author discusses "Accuracy of the analysing technique," which involves methods of counting grains under the microscope, grain count and accuracy, and the application of more complex statistical methods to heavy mineral study (an application that the author does not approve wholeheartedly).

Chapter III (25 pp.) on "Causes and Effects of Granular Variation" has to do chiefly with local variations in heavy mineral content—that is, variations within a mineral association, or even variations in samples taken in a traverse across a stream bed. This chapter is perhaps the most important one for those interested in methods and problems of stream transport, and the conclusions reached recur again and again in the rest of the report. Only a few of these conclusions can be covered here.

First, the author examines the relative importance of size, shape, and density in sedimentation under given current conditions. Shape he rules out as being of relatively slight importance. Density likewise, seems to have but slight effect in many cases. This is proved in part by comparing the size distribution of light fraction and a chosen heavy

mineral species in several samples of different mechanical composition and coefficient of sorting. Some heavy mineral species show size distributions very similar to that of the lights, others may be much smaller in size than the lights, but have a cumulative curve of similar shape and range of sizes; still other species may have a much more restricted range of size than the lights. Andel says that these variations do not check with the densities of the mineral species chosen; density, then, is not the principle factor. He believes that this factor is, in many cases, the original size of the mineral supplied to the sediment. "The relation between the size frequency distributions of the heavy minerals and that of the light fraction is in some cases distinct. Usually, however, the influence of the original size predominates to such an extent that no relation can be observed at all."

A further way of showing that size sorting is not necessarily the chief factor in determining size distribution of heavy minerals consists of taking (total) samples with almost identical size distributions (cumulative curves) and then analyzing each sample for heavy mineral content (weight) in each grade size. Samples with almost identical mechanical composition may have very different amounts of heavy minerals in the different grades (size fractions). This points to the conclusions that in the formation of two sediments of identical mechanical composition, different sizes of heavy minerals were available in the two cases. This is further borne out by the fact that in the same size fraction of one sample, hydraulically equivalent minerals may occur in quite different proportions.

A further consequence of the above is that "the supplied sands have a locally varying composition." Samples taken transverse to a river course show considerable variation; the sediment just downstream from this profile, and derived from it, will probably show a similarly great variation, since it is unlikely that much mixing or sorting will have occurred between these two nearby transverse profiles. That good mixing and sorting does not occur for long distances in the Rhine is shown by local "pockets" or "bars" of diagnostic minerals as much as several hundred kilometers below the mouth of the tributary yielding them. Such poor mixing would not be expected in sediments worked over by waves and currents.

Since there is all this variation, what size fractions should be studied in comparing different samples? Usage has differed, but the author shows that no one or few fractions are adequate, since much of the variation is not a function of size, but of availability. Because of the much greater labor involved in making size fraction analyses, the author recommends that they be reserved for solving special problems and he believes that most of the work can be accomplished with "normal analyses"—i.e. made without splitting the samples into their constituent size fractions.

Chapter IV, "Description of Minerals," (12 pp.) will probably be the one of least interest to those who will not have to deal with these minerals.

In general the author has attempted to describe only those features of importance for his particular varieties, but sometimes he includes data which are true for the species as a whole. For example, under enstatite, "extinction parallel to cleavage" and "length slow"; and under titanite, "high refractive index and birefringence" do not help us recognize any particular variety of these two minerals.

For the most part, varieties are defined by pleochroic colors and their optical framework. To the reviewer, verbal descriptions of colors seen under the petrographic microscope convey but little, but what else can one do? One important thing has already been done: Andel reports that there is a complete type collection of all minerals and mineral varieties available for comparison at Wageningen.

Chapter V, "Mineral Associations of the Supply Areas," (36 pp.) seeks to determine the source of the heavy mineral associations found in the Rhine sediments. The method of work was to collect samples (about 150 in number) from the modern stream sediments of all the important Rhine tributaries and from the important affluents of these tributaries. In some cases, terrace or hard-rock samples were used for supplementary information on heavy mineral sources. The location of all samples is shown on a map. In general, the problem was not to trace the minerals back to their ultimate source, but in cases where this information was already available in the literature, pertinent references are given.

The material is presented in considerable detail, with tables, longitudinal profiles showing variations in heavy minerals along certain tributaries, diagrams of mineral content in certain size fractions, and other helpful figures. From these data, the author distinguishes nine provinces which yield significantly different assemblages to the Rhine sediments; these provinces are likewise shown on a map. A few of these provinces, as presented in Andel's chapter summary, are given below:

"Aare province: main components are the alterites, accompanied by garnet, epidote and accessory blue-green hornblende. The Tertiary Molasse and partly also the Alps are the source of this homogeneous province.

"Moselle province: derived from two sources, the Vosgian crystalline rocks (angular garnet, blue-green and brown hornblende) and the Triassic (rounded garnet, tourmaline, zircon and rutile).

"A-province: principal minerals are epidote and garnet with accessory blue-green hornblende, in varieties that differ very little from those of the Aare-province. Glacial deposits of Fenno-scandian origin are the source of this group."

One point has rather confused the reviewer. In this chapter the nine units spoken of above are called "distributive provinces" by Andel, and he says about them: "... grouped into distributive provinces, comprising all river sands that constitute together a genetical, regional and chronological unity." When this is compared with the definition of a sedimentary petrological province: "... a group of sediments which constitute a natural unity by age, origin, and distribution," I am not sure what the difference is, unless one is thought of as an "area" of supply and the other of sedimentation.

Chapter VI, "Mineral Associations of the Rhine," (28 pp.) is a very interesting account which, using all the data previously described, plus the examination of some 150 samples from the Rhine itself, attempts to show how the various sources and the different agencies have made the Rhine sediments and Rhine mineral assemblages what they are today.

Again, the data are skillfully handled, with a large number of figures and diagrams. A longitudinal profile of the Rhine shows variation of mineral assemblages along its course, using "normal analyses." Another shows similar variation for the different size fractions. Individual mineral species are treated. For example, one figure shows progressive variation of zircon percentages along the River, and another gives progressive variations of the first and third quartiles, the median, the sorting and skewness of garnet for different parts of the Rhine.

Some of the important final conclusions may be enumerated: 1) That the Rhine sediments have quite different character and mineral assemblages in the upper part (Lake Constance to Bingen), middle part (Bingen to Bonn), and the lower course (below Bonn). The upper part has finer grained sediments, which show much less variability in mineral content. The middle part contributes a lot of coarser-grained sediment, with a large proportion of volcanic minerals. The lower part has sediments which result from a partial mixing of the two upstream types, plus sediment (a large proportion) derived from erosion of the bed

and banks; the sediments of this part show much greater variation in heavy mineral content. 2) "Usually supply by affluents only changes the composition of the main river sediments for a limited distance and in a comparatively small measure." 3) "Apparently transport and sedimentation in the Rhine are essentially limited to the local reworking and redepositing of old bottom sediments. Hence river transport of sand is an extremely slow process." 4) "The influence of an affluent on the heavy mineral association of the Rhine is not determined by the extent of its drainage area and often even not by the amount of sediment that is supplied by it." For example, some of the small streams in the Rhenish Schiefergebirge supply large proportions of heavy minerals in their small contribution of sediment; large rivers like the Main and Neckar may supply large quantities of light minerals but very few heavies. 5) "The abrasion of heavy minerals in a large river such as the Rhine is negligible." Finally, the author warns that his conclusions have been drawn from the Rhine only, and that his reader should be careful in applying them to other rivers or river systems.

A few general comments are perhaps in order. First, the report should be commended. It seems to have been well conceived and to have been carried out in a sound fashion. The method of presentation is excellent, and the English, though at times having a slight "foreign" flavor, never leaves any doubt as to the author's meaning. The one thing that seems lacking in a work of this size and importance is an index. Second, the reviewer, through lack of space, time, and ability, has not given a critical treatment of Andel's data. Particularly in the field of sorting and grain size, influence of availability on mineral composition, and in similar questions there may be interpretations different from those given by the author. A second paper, on "grain size and abrasion" is promised, and that may be the time for another reviewer to consider all his data from a rigorous statistical viewpoint. Third, why is it that we have no similar studies in the United States? It certainly is not true that we have no situations comparable to Holland and the Rhine, and we have no lack of opportunity to apply similar methods. Offhand, the reviewer remembers only two studies, by Russell and by Stowe, which are at all like Andel's report in purpose and method. It is certainly only a question of time before such interesting and important work is attempted in our country.

LINCOLN DRYDEN
Bryn Mawr College

Textbook of Petrology, vol. one, *The Petrology of the Igneous Rocks* by F. H. Hatch, A. K. Wells, and M. K. Wells, 469 pp. Thomas Murby and Company, London, 10th edition, 1949. Price $4.00.

This book was first published in 1891 and since that time it has undergone many changes and revisions, growing to nearly four times its original size. The earlier editions appeared every two or three years but the later editions were revised approximately every decade. These frequent changes were made in an attempt to improve quality and bring the contents up to date. The long life of this work indicates the degree of success with which it met as a text and reference for rock study. The tenth edition should prove no less popular and valuable.

It is believed that the first chapter should never have been incorporated in the text. This eleven-page treatment of optical properties serves no useful purpose, being entirely too brief to instruct the uninitiated in the theory of crystal optics and the function and use of the polarizing microscope. It does not even serve as an adequate review for those who were one time more experienced in microscopic work. The authors should assume that the reader either has or has not an understanding of the fundamentals of optical mineralogy; the treatment should be made complete or not at all.

The chapter on the description of the

rock forming minerals has been thoroughly revised, and this is one of the most admirable features of the book. For the first time the subject matter is based on atomic structure and a fairly successful attempt is made to interrelate structure, chemical composition, optical properties and crystallography. This is not only a more logical and correct approach but, in the long run, a simpler approach to the study of petrologic mineralogy. It is to be regretted that such a revision was not forthcoming in the ninth edition of this work and that the present edition is not more up to date in this respect. Nevertheless it is a great step forward and as a result the value of the text is greatly increased.

A brief but very interesting discussion of rock classification introduces the numerous chapters on rock description. The requirements and basis of classification are logically presented. The silica percentage of rocks, which was used partly as a basis of classification in all earlier editions, has been abandoned and the authors adhere strictly to mineral composition and grain size. An amazing amount of material is included in the rock descriptions and the numerous and large illustrations should prove very helpful to the beginning student.

Two short chapters deal with a few problems of petrogenesis; one of them briefly covers the problem of origin of granitic rocks. Though very brief indeed, this treatment should help to stimulate the student and direct his thinking along petrologic lines.

A good index is one of the most helpful and useful parts of any textbook. A rather hurried check, however, indicates that many terms used in the text, some of which are shown in italics and bold face type, are not included in the index. It is believed that, had the index of this edition been extended by several pages to include these terms and all rock names used as well, the text would serve as a splendid reference book.

CARLETON A. CHAPMAN
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Textbook of Petrology, vol. two, *The Petrology of the Sedimentary Rocks*, by F. H. Hatch and R. H. Rastall, 3rd edition revised by M. Black, 383 pp., 75 figs., Thomas Murby and Company, London, 1938, reprinted 1950.

This third edition of *The Petrology of the Sedimentary Rocks* was reviewed in the *Journal of Sedimentary Petrology* in December, 1938. The occasion for the present review is the issue of a reprint in 1950.

When this volume was published in 1938 it was an up-to-date though somewhat brief treatment of the field, and one which was recommended to every student of sediments. The viewpoint expressed on page 39—"The study of the sedimentary rocks is not yet sufficiently far advanced to make a strictly logical and consistent classification possible, and it is inadvisable to force the sediments prematurely into a rigid or elaborate classification; such a procedure would tend to obscure rather than to clarify the important issues"—is still refreshing today.

It is unfortunate that this excellent little book was not revised before reprinting. In view of the advances made in the field during the past twelve years, and particularly in view of the number of recently published books dealing with the same subject material, *The Petrology of the Sedimentary Rocks* must now be considered out of date.

J. L. HOUGH

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MARCH, 1951

ANNOUNCEMENT

CREATION OF AN ASSOCIATION OF FRENCH SEDIMENTOLOGISTS

Dr. Andre Vatan has sent to the *Journal* an announcement of the creation of the "Groupe de Sedimentation," an association of French sedimentologists which intends to have two or more meetings every year, generally on a determined subject. For the present, the association will have a mimeographed publication. General results will be presented in larger publications, such as Bulletin de la Société géologique and C. R. Acad. de Sciences.

Letters concerning this association are to be addressed to the general secretary, M. A. Vatan, Institut du Petrole, Rueil (S.O.), France. The constitution is reproduced herewith:

CONSTITUTION D'UNE ASSOCIATION POUR L'ETUDE DES ROCHES SEDIMENTAIRES ET DES PHENOMENES DE SEDIMENTATION

I. *Nom:* L'association portera le nom de:
GROUPE DE SEDIMENTATION

II. *Objet:* L'objet de l'association est l'étude en commun des roches sédimentaires et des phénomènes de sédimentation.

Ce but sera atteint par des réunions périodiques consacrées à des exposés et des discussions.

III. *Membres:* L'association est ouverte à tous les travailleurs intéressés par l'étude de la sédimentation et des roches sédimentaires. A titre d'invités, elle pourra recevoir des chercheurs des disciplines connexes: zoologie, paléontologie, océanographie, géographie physique etc. . . . pour des exposés se rapportant à ces disciplines dans leurs rapports avec la sédimentation.

IV. *Organisation:* L'association est dirigée par un conseil de 7 membres. Il n'y a pas de président—un secrétaire général, élu pour 4 ans, faisant partie du conseil, rédigera le procès-verbal de chaque réunion. Il sera charge de l'organisation générale et se tiendra en relations avec tous les membres de l'association pour leur communiquer tous les renseignements concernant l'association. Le conseil sera élu par l'assemblée générale et renouvelable par 1/3 chaque année. Le ler tiers renouvelable sera désigné au sort.

V. *Reunions:* L'association se réunira en principe deux fois par an sans que ce nombre soit impératif. Ces réunions se tiendront à tour de rôle dans chacun des laboratoires intéressés. Certaines pourront avoir lieu sur le terrain. Des réunions extraordinaires pourront avoir lieu à l'occasion du passage de visiteurs de marque.

Dans ces colloques seront exposés et discutés des sujets déterminés. Le sujet principal, choisi par la commission, sera approuvé par l'assemblée et annoncé au moins 6 mois à l'avance. La réunion sera dirigée par un président de séance.

VI. *Commission:* MM. Prof. BIROT, Prof. J. BOURCART, Mme Y. GUBLER, M. LUCAS, M. MILLOT, Prof. RIVIERE, secrétaire général A. VATAN.

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